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A Global Analysis of the Ordovician–Silurian boundary

Edited by L. R. M. Cocks & R. B. Rickards

Geology Vol 43 28 April 1988

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A Global Analysis of the Ordovician–Silurian boundary

Edited by **L. R. M. Cocks**
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The papers incorporated in this volume represent contributions from the International Working Group on the Ordovician–Silurian Boundary, a constituent body of the International Commission on Stratigraphy within the International Union of Geological Sciences.

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Introduction

L. R. M. Cocks & R. B. Rickards

The base of the Silurian System was agreed by the I.U.G.S. Executive Committee in May 1985 (published June 1985 in Bassett 1985), and was taken at the base of the *acuminatus* Zone at Dob's Linn, Scotland (Cocks 1985).

This volume closely reflects the achievements of the Ordovician–Silurian Boundary Working Group from its formation in 1974 to its disbandment in 1985. A detailed account of the activities of the Group is given in the next chapter, including the procedures followed which led to the decision on the definition of the boundary. We have taken the opportunity to gather in this book a global review of the Ordovician–Silurian boundary. These contributions are partly based on submissions on places and fossil groups made during the lifetime of the Working Groups and circulated by the Secretary, but these, if used in this volume, have been thoroughly updated by the respective authors and their colleagues. In addition we have commissioned a number of papers to give an overview of the many places where the boundary is exposed, as well as others on the global analysis of sedimentary events, and the evolutionary progress of the most important biological groups across the boundary.

It has always been clear from discussions that unanimous agreement would never be possible. Different countries have different traditions and philosophies, for example with respect to stratigraphical principles. This is especially true of the concepts of zones, and of the utility of zones for correlative purposes. For example, Mu (this volume) attempts a very detailed correlation of what are regarded elsewhere as potential subdivisions of the *acuminatus* Zone, claiming that an *ascensus* fauna underlies the *acuminatus* Zone (as it is, indeed, seen in China). But in some of the most precisely and exhaustively collected sections, such as at Dob's Linn, Scotland, it seems clear that the two species appear more or less simultaneously, albeit with *ascensus* more abundant low in the zone, and *acuminatus* more common in the upper part of the zone and outlasting *ascensus*. Thus, whilst there is a case for locally subdividing an *acuminatus* Zone, as Teller (1969) and others have sensibly done, it should be made clear that on current information these subdivisions correlate in total with the *acuminatus* Zone at Dob's Linn. In sections where the record is perhaps not very complete, or the fauna not abundant, it may appear that *acuminatus* follows *ascensus*.

Barnes (this volume) considers that, although the systemic boundary has now been fixed, its 'reconsideration may be necessary' (Lespérance *et al.* 1987). The main grounds for this opinion are that the Anticosti sequence has a future potential for further studies; has all the attributes for a boundary stratotype; and that 'important stratigraphic principles have been disregarded or overruled in making the final stratotype decision'. It cannot be overemphasized that the procedures adopted by the Working Party Group throughout its life were correct, proper, democratic, and always in accord with I.U.G.S. guidelines and with specific guidance from I.U.G.S.

If some stratigraphical ideas have been disregarded or overruled, then a substantial majority of the Working Group took the decisions to do so: the voting which took place is recorded in the next section. 'Potential' is always a difficult commodity to evaluate: and the judged potential of a section cannot delay for ever what will always be arbitrary decisions in the end. By the time a reconsideration was worked through (? ten years) another section would no doubt be vying with Anticosti in terms of its potential. Where then?

That Anticosti has most of the attributes necessary for a boundary stratotype is beyond question. That is why it was on a short list of two, voted upon by the Working Group. Other sections were of an almost equally high standard, for example, in China and the Lake District of England. But Anticosti does have one very serious drawback in any current discussions on

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correlations about the boundary, and that is its seemingly poor record of graptolites. It may be that at some future time graptolites may be relatively demoted in value for correlative purposes, but that time is still far away on present information. Dob's Linn *also* has most of the attributes of a boundary stratotype, and the Working Group, after eleven years of study, considered it better than Anticosti. In fact, the boundary has now been certainly put at the correct level, using the best group for correlation, the graptolites. Despite the fact that the *Hirnantia* brachiopod fauna is very often overlain by *persculptus* Zone graptolites, unequivocal evidence from both Kazakhstan (Koren *et al.* this volume) and the Lake District of England (Cocks this volume) shows that it also occurs rarely within the *persculptus* Zone. There is a strong feeling amongst most biostratigraphers that they prefer to regard the *Hirnantia* fauna as Ordovician rather than Silurian in age and not straddling the systematic boundary, and this assignment to the Ordovician can be achieved only by a sub-*acuminatus* Zone boundary, as was eventually decided.

A more interesting question is the precise age, in terms of graptolite zones, of the maximum glacio-eustatic drop in sea level, and this is still not yet definitively answered although it was probably about half way through the *persculptus* Zone—there are some well-dated *persculptus* bearing post-glacial transgressive beds in parts of North Africa. On the other hand, the precise duration and extent of the glacial episode (Fig. 1) certainly varied from place to place—commencing even in late Caradoc and early Ashgill times in some parts of Gondwana, and certainly continuing into post-*Hirnantia* fauna times, perhaps into the Rhuddanian, in others, e.g. South Africa. It is also important to note that detailed investigation indicates that the 'end

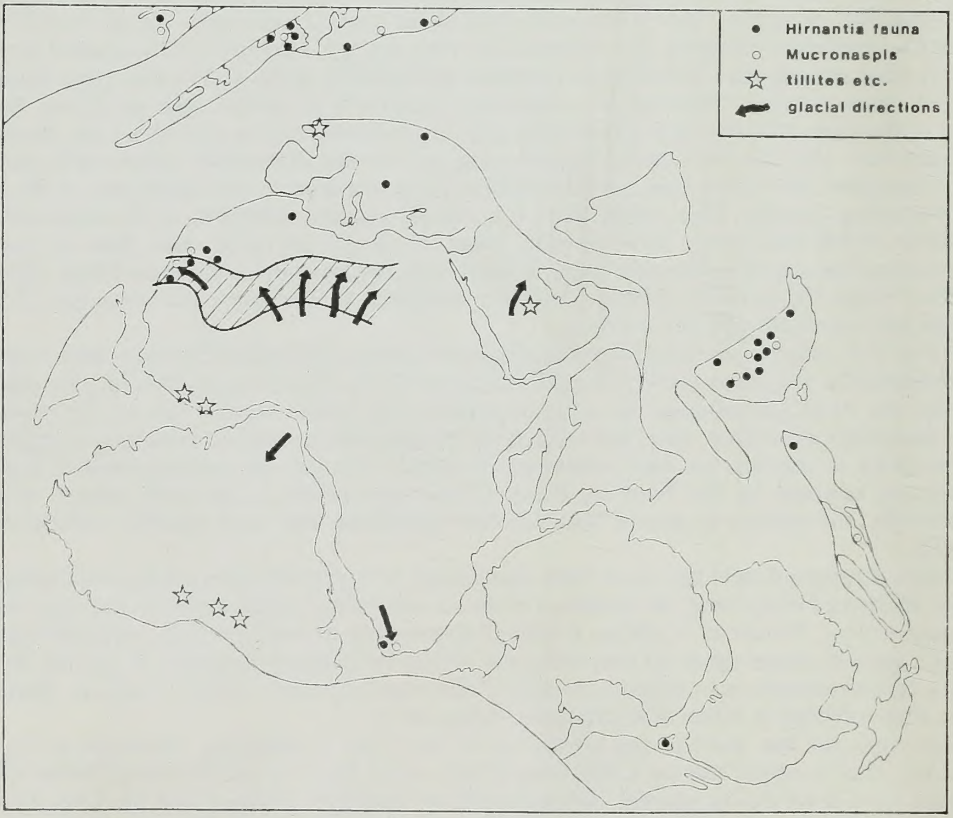

Fig. 1 Distribution of the latest Ordovician glacial deposits in Gondwana and adjacent areas (after Cocks & Fortey 1988).

Ordovician' faunal extinctions were by no means synchronous. No other faunal or floral group than graptolites yet approaches the sensitivity and exactness of the graptolites during the period in question—for example from the mid-Ashgill (base of the Rawtheyan) to the end of the early Llandovery (Rhuddanian) there are no fewer than eight graptolite zones, as compared with three or four conodont zones, and four successive brachiopod faunas, three or four ostacod faunas, three or four trilobite faunas etc. This, from a period of only perhaps 7 or 8 million years (McKerrow *et al.* 1985), makes the graptolites compare well with Mesozoic ammonites or Tertiary foraminifera as a precise dating tool.

The coverage in this volume of the Ordovician–Silurian sections themselves cannot be total partly because several regions are little known. However, it is worth drawing attention here to probable additional Ordovician–Silurian boundary sections in Libya (Klitzsch 1981), Burma (Mitchell *et al.* 1977; Wolfart *et al.* 1984) and Greenland (e.g. Hurst & Kerr 1982; Surlyk & Hurst 1984). In addition we are aware of preliminary work on strata about the boundary in Vietnam, Thailand, Malaysia and other parts of SE Asia. In the instance of central Nevada, U.S.A., we have not republished a revised preliminary submission because there is nothing yet new to add to the work by Berry (1986). There is also further work in preparation on Scandinavia.

We would like to end this introduction with a tribute to the many people involved, both as members of the Working Group and as contributors to the present volume, who patiently took part in the meetings, newsletter, activities and final decision-making, and thank them all for their patience, support, good humour and international friendship; despite the controversy of the eventual scientific conclusion.

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The Ordovician–Silurian Boundary and its Working Group

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Synopsis

After a brief history of the study and definition of the Ordovician–Silurian boundary in the nineteenth and early twentieth centuries, the process of setting up the Ordovician–Silurian Boundary Working Group is described, together with its progress, publications and final decision-making during the period 1974–1985.

Both the Cambrian and the Silurian Systems were established as formal system names by Sedgwick and Murchison respectively amicably enough in 1835, but during the next thirty years it become clear that the upper part of Sedgwick's Cambrian occupied the same time and space as the lower part of Murchison's Silurian. It was not until after the deaths of both men that Charles Lapworth in 1879 established the Ordovician System to occupy the chief overlapping ground between the older part of the Silurian and the younger part of the Cambrian. In contrast to the rather generalized earlier definitions of the boundaries of the Cambrian and Silurian, Lapworth's definition of the limits of the Ordovician was admirably precise: he defined the new Ordovician System as the 'strata included between the base of the Lower Llandovery formation and that of the Lower Arenig' (Lapworth 1879: 14). There were subsequently problems (which are still not entirely resolved today) about the position and international correlation of the 'base of the Lower Arenig', but these are the province of the Cambro–Ordovician Boundary Working Group and will not be further discussed here. 'The base of the Lower Llandovery' has been much less ambiguous, and thus in general any dispute surrounding the definition of the Ordovician–Silurian boundary has always been of a much lesser magnitude than the problems of the Cambro–Ordovician and the Siluro–Devonian boundaries.

From the time of Murchison onward, 'the base of the Lower Llandovery' was defined primarily in terms of shelly facies and without much precision, and usually recognized by the incoming of various pentameride brachiopods such as *Stricklandia*. However, following Charles Lapworth's classic work on the Ordovician and Silurian rocks of Scotland in the period 1870 to 1880, it became clear that the best national and international correlation tool in rocks of those ages was the sequence of graptolite zones, and these zones were subsequently used in practice, with Lapworth himself, and subsequently the great graptolite monograph of Elles & Wood (1901–1918), using the *acuminatus* Zone (type locality Dob's Linn, Scotland) as the *de facto* base of the Silurian. The *acuminatus* Zone was poorly developed as such in Wales, and so Jones (1909) erected the *persculptus* Zone (type locality Pont Erwyd, Wales), which was subsequently realised to be of the same age as the lower part of Lapworth's broad *acuminatus* Zone in Scotland. Thereafter most stratigraphers treated the *persculptus* Zone as the base of the Silurian, e.g. in the *Lexique stratigraphique international* (Whittard 1961), and this horizon was also taken as the base of the Silurian by Cocks *et al.* (1970) when they erected stages for the Llandovery Series, with a basal boundary defined at Dob's Linn.

It was probably the identification of the problems surrounding the Silurian–Devonian boundary and their subsequent illumination and solution that gave impetus to the international effort and will to define properly the exact horizon and identify a type locality for the various systemic divisions of the Phanerozoic. The Siluro–Devonian Boundary Working Group worked formally between 1960 and 1972 (Martinsson 1977), but that work was preceded by a period of uncertainty, during which some of the procedures within the International Geological Congresses and the International Union of Geological Sciences were being developed.

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And so it was during the Ordovician–Silurian symposium at Brest, France, in 1971 that Claude Babin was the first to identify vocally the need for a group to be formally established to investigate and stabilize the Ordovician–Silurian boundary. This was put to the nascent Commission on Stratigraphy at the International Geological Congress in Montreal, Canada, in 1972, who felt that such a boundary working group should be established not by that commission directly, but at a suitable international meeting and through the joint coordination of the then proposed Ordovician and Silurian Subcommissions. These last two bodies were finally established at the Ordovician Symposium at Birmingham, England, in September 1974, and one of their first acts was to arrange the initial meeting of the Ordovician–Silurian Boundary Working Group, which first met at Birmingham on 19th September 1974. Those present at that meeting were C. Babin (France), C. R. Barnes (Canada), S. M. Bergström (USA), A. J. Boucot (USA), L. R. M. Cocks (UK), J. Destombes (Morocco), J. K. Ingham (UK), V. Jaanusson (Sweden), P. J. Lespérance (Canada), D. J. McLaren (Canada), L. Marek (Czechoslovakia), F. Martin (Belgium), R. B. Rickards (UK), P. Sartenaer (Belgium), N. Spjeldnaes (Denmark), L. Teller (Poland), J. T. Temple (UK) and E. A. Yolkin (USSR). It was decided that 6 voting members of the Working Group should be nominated by both the Ordovician and the Silurian Subcommissions, plus their two chairmen *ex officio*, and that 3 voting members from the USSR and 1 from Czechoslovakia should be nominated by their respective academies of science. Thus the Ordovician Subcommission nominated Barnes, Bergström, W. B. N. Berry (USA), Destombes, Ingham and Jaanusson, with A. Williams (UK) *ex officio* as their Chairman, and the Silurian Subcommission nominated Boucot, Cocks, S. Laufeld (Sweden), Lespérance, Rickards and Temple, with Spjeldnaes *ex officio* as their Chairman. Any interested and active worker on Ordovician–Silurian boundary problems could be accepted as a Corresponding Member. At that first meeting R. B. Rickards was elected by those present as the Chairman of the Working Group, and L. R. M. Cocks as the Secretary. It was also decided that most of the Group's activities and communication would take the form of circulars to be issued by the Secretary, and this is what subsequently happened, although field and discussion meetings also took place, and that the circulars should include reports on various Ordovician–Silurian sections or countries and also on the different fossil groups. The first circular was issued in October 1974: it reported the formation of the Working Group, and listed which members had promised to prepare reports.

In the next few years many circulars were issued, which included reports on boundary sections in Australia, Austria, Belgium, Canada (many areas), China, Czechoslovakia, England, France, Italy, Morocco, Poland, Scotland, Sweden, Wales, USA and USSR (Altai Mountains, East Baltic, Kazakhstan and NE Siberia), and also on acritarchs, chitinozoa, conodonts, graptolites and physical changes near the boundary. Many people became Corresponding Members, and the Voting Members were increased by D. L. Kaljo, T. N. Koren and I. F. Nikitin from the USSR, L. Marek from Czechoslovakia and Mu En-zhi from China, all of these nominations being accepted and ratified at the appropriate times by the I.U.G.S. Commission on Stratigraphy, the parent body of the Working Group. Meetings were held at the International Geological Congress at Sydney, Australia, in August 1976 and informal meetings at Alma-Ata, USSR, in May 1977 and at the Ordovician Symposium at Columbus, USA, in August 1977, and it became clear that a more substantial meeting of the Working Group would be valuable so that future plans of action could be formulated. This coincided with an expressed wish by various geologists to see the classic sections of Great Britain, and accordingly a meeting was arranged from 30th March to 11th April 1979, jointly with the Silurian Subcommission. By that time R. J. Ross jr and C. H. Holland had taken over the chairmanships of the Ordovician and Silurian Subcommissions respectively.

Those attending the British meeting in 1979 were (Voting Members of the Ordovician–Silurian Boundary Working Group with an asterisk): *C. R. Barnes (Canada), M. G. Bassett (UK), *L. R. M. Cocks (UK), *C. H. Holland (Ireland), *J. K. Ingham (UK), J. S. Jell (Australia), Jin Chun-tai (China), *D. L. Kaljo (USSR), P. Legrand (France), *P. J. Lespérance (Canada), Lin Bao-yu (China), F. Martin (Belgium), A. Martinsson (Sweden), *Mu En-zhi (China), *R. B. Rickards (UK), H.-P. Schönlaub (Austria), B. S. Sokolov (USSR), L. Teller

Fig. 1 The British field meeting, April 1979, outside Ludlow Castle, Shropshire. From left to right L. R. M. Cocks, Jin Chun-tai (obscured), B. D. Webby, C. R. Barnes, J. S. Jell, Wang Wei, Lin Bao-yu (obscured), D. Kaljo, Mu En-zhi, D. J. Siveter, F. Martin (obscured), L. Teller, P. J. Lespérance (obscured), D. E. White, A. Martinsson, B. S. Sokolov, P. Legrand, J. T. Temple, H. P. Schönlaub, M. G. Bassett, R. B. Rickards. (Photo C. H. Holland).

(Poland), *J. T. Temple (UK), G. B. Vai (Italy) and B. D. Webby (Australia). Thus more than half the Voting Members and a considerable breadth of both stratigraphical and palaeontological expertise were represented (Fig. 1). Sections were examined in Wales (Llandovery, Meifod, Hirnant and Pont Erwyd), the Lake District of England (Yewdale, Skelgill and Spengill), and Scotland (Dob's Linn), but, more importantly, business meetings were held in the evenings. Following a long-standing tradition of the Commission on Stratigraphy (whose then Chairman, Martinsson, and Secretary, Bassett, were present) all of the people present were allowed to participate freely in the discussions and also to take part in the informal voting which took place.

The various animal and plant groups were discussed and reviewed in turn, and it was agreed that only graptolites, brachiopods, conodonts, and to a lesser extent trilobites, were important in the Ordovician–Silurian boundary discussions in the present state of knowledge. Localities were then considered. Having inspected the type Llandovery area, all members present were unanimous in rejecting that area as the boundary stratotype, large due to the unfossiliferous nature of the A_1 Sandstone of Jones (1928) at the base of the succession, the sporadic exposure near the base, and the lack of stratigraphically critical fossils, particularly graptolites and conodonts, then known from beds near the boundary (although this situation has been much improved by subsequent work, Cocks *et al.* 1984). Other localities were graded in turn, with the following scheme: A, a possible section for placing the boundary; B, an important section which may be considered further in discussing the boundary, and C, a section or area unlikely to prove important in boundary definition. The grading was as follows:

A Anticosti Island (Canada), Dob's Linn (Scotland).
B Carnic Alps (Austria), Cornwallis Island (Canada), Hupei (China), Mirnyi Creek (Siberia,

USSR), Missouri (USA), Nevada (USA), Pont Erwyd (Wales), Szechuan (China) and Yewdale Beck (Lake District, England).

C Australia, Bala district (including Hirnant area, Wales), Belgium, Bohemia (Czechoslovakia), France, Garth (Wales), Hudson Platform (Canada), Kazakhstan (USSR), Kweichow (China), Lake District (apart from Yewdale Beck, England), Manitoba (Canada), Manitoulin Island (Canada), Morocco, Newfoundland (Canada), North American mid-continent (except Missouri and Nevada), Pembrokeshire (Wales), Percé (Canada), Poland, Scania (Sweden), Shensi (China) and Yukon (Canada).

In addition the Working Group then felt that more reports were needed from Algeria, Bornholm (Sweden), Burma, Dalarna and Västergötland (Sweden), Estonia (USSR), India and the Himalayas, Norway, Rae Grain (Scotland), Portugal, South America, Spain and West Nevada: however, although more data on some of these areas were subsequently gathered, none proved to have much extra to offer in the main definition of a stratotype. Because Anticosti Island, Canada, was one of the leading contenders for the definitive boundary section, it was agreed that a further field meeting should be held there. Other briefer meetings were also held in Paris, France, during the 1980 International Geological Congress, and in the Carnic Alps of Austria in late July and early August 1980. Meanwhile the debate persisted as to the best method of correlation across the boundary interval, and whether the actual boundary should be defined by the use of conodonts or graptolites. It was generally agreed that brachiopods and trilobites should not be used in the definition, except that there was a strong feeling that the widespread *Hirnantia* brachiopod fauna should be included within the Ordovician rather than the Silurian.

The Working Group circulars also contained various discussion and position papers between 1978 and 1982 on the theory and practice of defining the boundary both geographically and biostatigraphically. Opinions differed as to whether or not the stratotype could be satisfactorily placed within a nearly exclusively graptolite sequence such as Dob's Linn, and, if the boundary was to be defined on graptolites, whether it was to be at the base of the *extraordinarius*, the *persculptus* or the *acuminatus* Zone. There was no real consensus on the answers to these questions.

The field meeting to Quebec, which was partly in Anticosti Island and partly in the Gaspé Peninsula, was held in July 1981, again jointly with the Silurian Subcommission. Those attending (apart from various other Canadian hosts) were T. W. Amsden (USA), *C. R. Barnes (Canada), *A. J. Boucot (USA), *L. R. M. Cocks (UK), *C. H. Holland (Ireland), P. Legrand (France), *P. J. Lespérance (Canada), F. Martin (Belgium), G. M. Philip (Australia), *R. J. Ross jr (USA), H.-P. Schönlaub (Austria) and L. Teller (Poland). This was a rather disappointing attendance, particularly of Voting Members, and hence the evening discussion meetings were not as representative of the differing positions of the complete group as they might have been if the attendance had been better. A review was given of each of the relevant biological groups, and general discussions ensued, with the following points noted. There were very favourable general impressions of the simplicity of structure and good exposure at Anticosti, but reservations on the lack of graptolites there near the Ordovician–Silurian boundary and the relative lack of work done on groups other than conodonts on the beds near the boundary. Opinions differed about the accessibility of Anticosti Island and also about the importance of the structural complexity of the Dob's Linn area. At the end of the meeting, two straw votes indicated that those present thought that Anticosti was the best available section across the Ordovician–Silurian boundary in the shelly facies, and that, other things being equal, it would be preferable to have the Ordovocian–Silurian boundary stratotype in the same area as the stratotype area for the lowest series of the Silurian System. The latter point was relevant since at that time Anticosti was one of the three candidates under consideration by the Silurian Subcommission (the other two being Llandovery itself and the Oslo Region, Norway) for the stratotype for the lowest Silurian series. Shortly after this meeting, R. B. Rickards resigned as Chairman of the Working Group, and, because it was clear that the decisions on the boundary were close to being taken, the Commission on Stratigraphy subsequently appointed the Chairmen of the

Ordovician and Silurian Subcommissions, R. J. Ross jr and C. H. Holland, as Co-Chairmen of the Group; which they remained until its closure.

After the formal circulation of a number of further views on the position and correlation of the future boundary stratotype through the Circular, and informal discussion between interested people, it was agreed that maximum publicity and attendance should be sought for a meeting of the Working Group at the Ordovician Symposium at Oslo, Norway, so that progress would be made on the boundary decision. At that symposium, two meetings of the boundary Working Group were held, as well as seven papers on the boundary being presented within the normal symposium sessions. The meetings, on 20th and 23rd August 1982, attracted 53 and 76 people respectively, including the following Voting Members: Barnes, Bergström, Berry, Cocks, Destombes, Holland, Jaanusson, Kaljo, Lespérance, Rickards and Ross. After lengthy discussion, the first decision taken was whether or not the time was yet ripe for a formal vote on deciding the boundary stratotype and horizons, and, despite strong pleas for delays to enable more research to be done from several speakers, it was decided by 47 votes to 14 that the time had now come. The choice of stratotype boundary had been narrowed to three:

(i) the first appearance of the conodont *Ozarkodina oldhamensis* at 50 cms above the Oncolitic Platform Bed at Ellis Bay, Anticosti Island, Canada.
(ii) the base of the *persculptus* graptolite Zone at Dob's Linn, near Moffat, Scotland.
(iii) the base of the *acuminatus* graptolite Zone at Dob's Linn.

At the Oslo meeting two informal votes were then taken: (i) Anticosti was preferred to the *persculptus* Zone at Dob's Linn by 34 votes to 13, with 25 abstentions; (ii) Anticosti was preferred to the *acuminatus* Zone at Dob's Linn by 35 votes to 13, with 26 abstentions. The same questions were also informally voted upon by the 30 Voting and Corresponding Members of the Working Group who were present, and 17 preferred Anticosti against 7 for the *persculptus* Zone (6 abstentions); and 19 preferred Anticosti against 5 for the *acuminatus* Zone (6 abstentions). Therefore, it was clear that a substantial majority of those at the meeting then preferred to place the base of the Silurian at Anticosti Island using conodonts, and that the Voting Members of the Working Group should take part in a formal postal ballot in the light of this knowledge. Thus Circular No 17 was distributed to the members in October 1982 with a ballot paper to be returned by the end of January 1983. There followed a period during which various letters were informally circulated and lobbying took place, although none formally through the Secretary apart from a paper by P. Legrand which was very critical of the Oslo decision and which was distributed with Circular 17.

At the end of the formal voting period, the votes returned stood as follows:

(i) Which do you prefer—Anticosti or the *persculptus* Zone at Dob's Linn?
For Anticosti: Barnes, Bergström, Boucot, Holland, Lespérance, Ross: total 6.
For *persculptus* Zone: Berry, Cocks, Destombes, Ingham, Kaljo, Koren, Laufeld, Marek, Nikitin, Rickards, Temple: total 11.
No vote received: Jaanusson, Mu: total 2.
(ii) Which do you prefer—Anticosti or the *acuminatus* Zone at Dob's Linn?
The votes received were identical to the *persculptus* Zone vote.

These results were distributed to all members of the Working Group in Circular 18 in March 1983. Since there had been an outright majority on the selection of Dob's Linn rather than Anticosti, this was accepted by the officers as a decision, and a second formal postal vote was called for, firstly to give Voting Members an opportunity to change their minds, and secondly to decide between the *persculptus* and the *acuminatus* Zones at Dob's Linn for the stratotype horizon. Opportunity was also given to the Corresponding Members to formally express their opinions. The results of this second ballot was announced in Circular No. 19 in August 1983, and were as follows:

(i) the place of the stratotype.
Voting Members. Dob's Linn: Berry, Cocks, Destombes, Holland, Ingham, Kaljo, Koren,

Laufeld, Marek, Nikitin, Rickards, Temple: total 12. Anticosti: Barnes, Bergström, Boucot, Lespérance, Ross: total 5. Abstain: Jaanusson, Mu: total 2. In addition 14 Corresponding Members voted for Dob's Linn, 8 for Anticosti, and 4 abstained.

(ii) the horizon of the stratotype.

Voting Members. Base of *acuminatus* Zone: Cocks, Holland, Ingham, Jaanusson, Kaljo, Koren, Marek, Nikitin, Rickards, Temple: total 10. Base of *persculptus* Zone: Berry, Destombes, Laufeld, Mu, Ross: total 5. Abstain: Barnes, Bergström, Boucot, Lespérance: total 4. 13 Corresponding Members voted for the base of the *acuminatus* Zone, 9 for the base of the *persculptus* Zone, and 5 abstained.

In addition the question of possible parastratotypes was also voted upon, with the possibility of erecting one parastratotype on Anticosti Island and the other in China, but on this question only 8 Voting Members voted for the erection of these, with 3 against and 8 abstentions, and so the officers decided not to proceed further on that topic, and they were assisted in that decision by informal advice against parastratotypes from the Commission on Stratigraphy.

Thus since there was a clear majority for placing the Ordovician–Silurian stratotype boundary at the base of the *acuminatus* graptolite Zone at Dob's Linn, Scotland, this decision was formally forwarded to the Commission on Stratigraphy for consideration with various other matters at their meeting at the International Geological Congress at Moscow, USSR in August 1984. The decision was endorsed by a postal vote of that committee, who subsequently forwarded it to the I.U.G.S. for ratification. The proposals were reported to a meeting of the full I.U.G.S. Executive Committee in Rabat, Morocco, in February 1985 and submitted to the I.U.G.S. Executive for a postal ballot, whose result was declared in May 1985, and published in June 1985 (Bassett 1985), together with an article describing the Ordovician–Silurian boundary at Dob's Linn (Cocks 1985). The Ordovician–Silurian Boundary Working Group was finally dissolved in its Circular No. 20, distributed in June 1985.

The life of the Ordovician–Silurian Boundary Working Group was therefore somewhat over ten years long, but it was useful not only in determining the position and horizon of the boundary itself, but also in stimulating a great deal of research in various parts of the world, and in encouraging international understanding and cooperation.

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Appendix

MEMBERSHIP OF THE ORDOVICIAN–SILURIAN BOUNDARY WORKING GROUP

Those names with an asterisk* were Voting Members, the remainder were Corresponding Members.

Amsden, T. W.	USA
Apollonov, M. K.	USSR
Babin, C.	France
*Barnes, C. R.	Canada
Bassett, M. G.	UK
Bergström, J.	Sweden
*Bergström, S.	USA
*Berry, W. B. N.	USA
Bolton, T. E.	Canada
*Boucot, A. J.	USA
Brenchley, P. J.	UK
Bruton, D. L.	Norway
*Cocks, L. R. M.	UK
Cramer, F. H.	Spain
*Destombes, J.	Morocco
Hamada, T.	Japan
*Holland, C. H.	Ireland
*Ingham, J. K.	UK
*Jaanusson, V.	Sweden
Jackson, D. E.	UK
Jaeger, H.	East Germany
Jin Chun-tai	China
*Kaljo, D.	USSR
Kobayashi, T.	Japan
*Koren, T. N.	USSR
*Laufeld, S.	Sweden
Legrand, P.	France
Lenz, A. C.	Canada
*Lespérance, P. J.	Canada
Lin Bao-yu	China
*Marek, L.	Czechoslovakia
Martin, F.	Belgium
Martinsson, A.	Sweden
McLaren, D. J.	Canada
*Mu En-zhi	China
*Nikitin, I. F.	USSR
Norford, B. S.	Canada
Nowlan, G. S.	Canada
Oradovskaya, M. M.	USSR
Poulsen, V.	Denmark
*Rickards, R. B.	UK
Rong Jia-yu	China
*Ross, R. J. jr	USA
Sartenaer, P. J. M. J.	Belgium
Schönlaub, H. P.	Austria
Sheehan, P. M.	USA
Sokolov, B. S.	USSR
Spjeldnaes, N.	Denmark
Teller, L.	Poland
*Temple, J. T.	UK
Toghill, P.	UK
Wang Xiao-feng	China
Webby, B. D.	Australia
Williams, A.	UK
Williams, S. H.	UK
Wright, A. D.	UK
Yolkin, E. A.	USSR

Dob's Linn – the Ordovician–Silurian Boundary Stratotype

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Synopsis

Dob's Linn, north-east of Moffat, southern Scotland, has been designated the Ordovician–Silurian boundary stratotype by the Ordovician–Silurian Boundary Working Group of the I.U.G.S. Commission on Stratigraphy. The boundary is placed at the base of the *P. acuminatus* Zone, marked by the first occurrence of *Akidograptus ascensus* and *Parakidograptus acuminatus*, *s.l.*, 1·6 m above the base of the Birkhill Shale in the Linn Branch section.

The stratigraphical interval covering this boundary consists of richly graptolitic black shale. Occasional metabentonites are also present. The underlying Upper Hartfell Shale is composed predominantly of pale grey-green, non-graptolitic shale and mudstone, with several black graptolitic bands referred to the *Complanatus*, *Anceps* and *Extraordinarius* Bands. The rich faunal assemblage of the *Anceps* Band reduces to only three diplograptid taxa in the *Extraordinarius* Band. This major extinction is recorded at an equivalent horizon by all other graptolitic sequences throughout the globe.

Sediments of the Upper Ordovician to Lower Silurian Moffat Shale Group were probably deposited entirely by distal turbidites in the abyssal depths of the Iapetus Ocean. Northerly-directed subduction subsequently transported the site of shale deposition into a proximal turbidite environment, resulting in a diachronous transition into coarse clastics of the overlying Gala Greywacke Group. Deformation related to subduction also produced imbricate thrusting and raised the area to a prehnite-pumpellyite facies metamorphic grade. Geophysical evidence indicates that the region is underlain by continental basement; this suggests that the Southern Uplands are allochthonous.

Historical introduction

Graptolites were first recorded from Dob's Linn, southern Scotland over one hundred years ago. The earliest publications to include descriptions of the fauna from the Moffat Shales (e.g. Carruthers 1858; Nicholson 1867; Dairon 1869; Hopkinson 1871) paid little or no attention to their stratigraphical importance. Elsewhere during this period, an ever increasing volume of articles on graptolites was being published, including a number which recognized their great potential for both regional and global correlation (Nicholson 1876). These included studies on the Lower Ordovician of northern England (e.g. Nicholson 1870, 1875), South Wales (Hopkinson & Lapworth 1875) and eastern Canada (Hall 1858, 1865; Billings 1865).

In 1864 Charles Lapworth obtained a teaching post connected with the Episcopal Church at Galashiels some 30 km north-east of Dob's Linn (Gibson 1921). He had no previous geological training or experience, but soon developed an interest in the local geology of the Southern Uplands. Harkness (1851) had described the repeated, faulted nature of this area composed of thick greywacke and containing a shale sequence termed the 'Moffat Series'. Otherwise this structurally complex region still defied satisfactory interpretation despite attempts by several other eminent geologists (e.g. Sedgwick 1850).

Lapworth's first publication on the Lower Palaeozoic (1870) concerned the geology of the Galashiels area. During these early years in his geological career, he recorded graptolites both from within the thick greywacke sequence of the Southern Uplands and from the underlying black shales. A summary of Lapworth's early lithostratigraphical division was published in 1872. During the next five years he completed an exercise of detailed geological mapping, logging of sections and bed-by-bed faunal collecting throughout the Moffat area. A selection of new graptolite taxa were figured and discussed briefly in 1876, while similar faunas were also illustrated from equivalent strata in northern Ireland (Lapworth 1877).

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Lapworth's major stratigraphical synthesis 'On the Moffat Series' was published in 1878, where he established beyond doubt the precise, ordered stratigraphical change in graptolite assemblages through the sequence of black and grey shales. With the exception of the Glenkiln Shale (best developed at Glenkiln Burn, south-east of Moffat) and the lowermost portion of the Lower Hartfell Shale (best exposed at Hartfell Spa, north of Moffat), Lapworth used the Main Cliff and Linn Branch sections of Dob's Linn as the standard reference for the Moffat Shale. While working at Dob's Linn, Lapworth stayed at Birkhill Cottage only a few hundred metres above the locality. The uppermost black shale division of the group was named after this cottage.

Lithological sections measured at Dob's Linn, together with graptolite assemblages and biostratigraphical divisions, were figured by Lapworth (1878: figs 27–30) in his major work, where the lithostratigraphical division of the Moffat Shale was also clearly defined. An earlier, less detailed log of the Moffat Shale from Lapworth's notes, covering the Glenkiln Shale to basal Birkhill Shale, is still preserved in Birmingham University, and is here illustrated for comparison (Fig. 1). Note that Lapworth's assignment of the lower part of the Upper Hartfell Shale to the 'Belcraig Shale' (after Beldcraig Burn near Moffat) was apparently never published.

During this research, Lapworth was appointed in 1875 to an Assistant Mastership at Madras College, St Andrews. In 1881 he was elected to the Chair of Geology at the recently established Mason College, Birmingham, which subsequently became the University of Birmingham. In addition to elucidating the structure of the Southern Uplands, Lapworth established the Ordovician System in 1879, solving the embittered feud between the schools of Murchison and Sedgwick (see Bassett 1985). He also made an equally painstaking, detailed stratigraphical study at Girvan on the south-west coast of Scotland (Lapworth 1882). His conclusions regard-

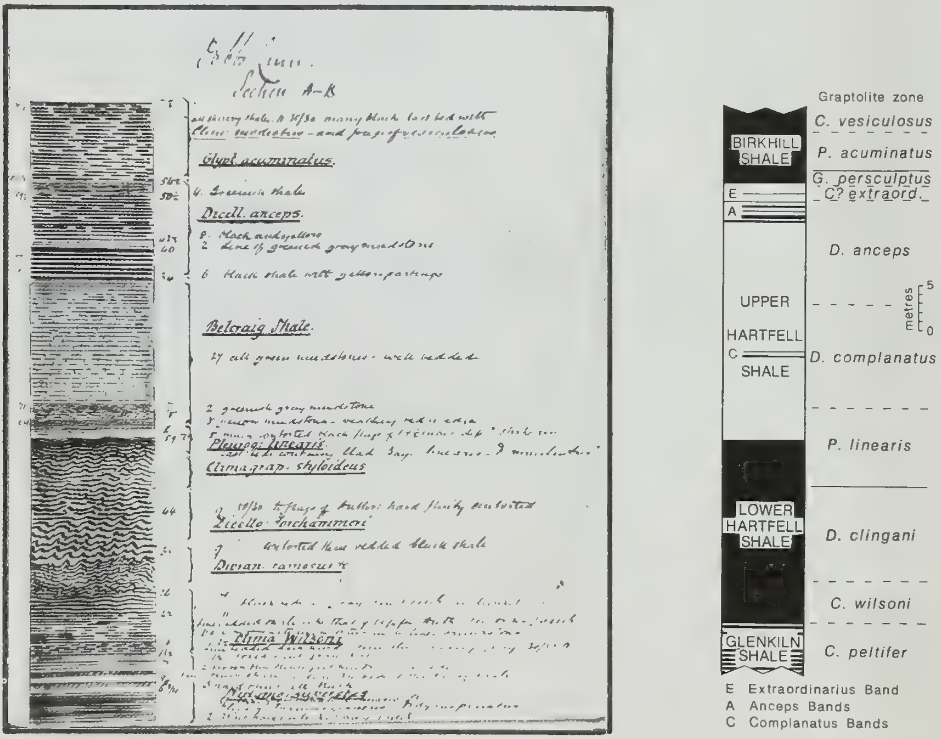
Section A–B
Glypt. acuminatus.
Dicell. anceps.
Belcraig Shale.
Pleurogr. linearis.
Climacograp. styloideus
Dicello. forchammeri
Dicran. ramosus
Climac. wilsoni
Graptolite zone
C. vesiculosus
P. acuminatus
G. persculptus
C? extraord.
D. anceps
D. complanatus
P. linearis
D. clingani
C. wilsoni
C. peltifer
BIRKHILL SHALE
E
A
UPPER HARTFELL SHALE
C
LOWER HARTFELL SHALE
GLENKILN SHALE
metres
5
0
E Extraordinarius Band
A Anceps Bands
C Complanatus Bands

Fig. 1 Reproduction of an unpublished section through Moffat Shale from Lapworth's notebook (preserved at Birmingham University), predating his modified version published in 1878. Modern measured section with graptolite zones is included for comparison.

ing the relationship between the strata at Girvan and those of the Southern Uplands were published in 1889.

The work of Lapworth was drawn upon heavily by Peach & Horne (1899), who also described many confirmatory sections through the Moffat Shale in the Southern Uplands. With the exception of one taxonomic paper published by Lapworth in 1880, most of his new graptolite taxa were first described fully by Elles & Wood (1901–18), whose work he supervised throughout its production. Following this major publication, little taxonomic or stratigraphical work was attempted at Dob's Linn for half a century. One notable exception was an article by Davies (1929), who included Dob's Linn in his detailed study of late Ordovician and early Silurian graptolites.

A series of recent biostratigraphical and taxonomic papers was initiated by Packham (1962), who described the evolution of *Glyptograptus tamariscus* and related diplograptids from the Birkhill Shale of Dob's Linn and from the Lower Silurian of the Rheidol Gorge, mid Wales. Toghill (1968) discussed the evolution of the earliest monograptids and formally established the presence of the *G. persculptus* Zone at Dob's Linn. He also gave a biostratigraphical summary of the entire Birkhill Shale with listings of the zonal assemblages (1968a), but none of the fauna was described or illustrated.

Toghill (1970) subsequently published a revision of graptolites from the Upper Hartfell Shale and top Lower Hartfell Shale. Brief taxonomic descriptions and illustrations were included; this paper added little in terms of refinement to Lapworth's biostratigraphical divisions, but was important in demonstrating the presence of previously unrecorded graptolite species. Cocks *et al.* (1970) used this data to make a premature proposal of Dob's Linn as the Ordovician–Silurian boundary stratotype, where they placed the boundary at the base of the Birkhill Shale (= 'base' of *G. persculptus* Zone). Rickards (1979) added the record of the *C.? extraordinarius* Zone, based on the discovery by Ingham (1979) of a black, graptolitic shale band midway between the top *Anceps* Band of the Upper Hartfell Shale and the basal Birkhill Shale.

A geological locality map of Dob's Linn was given by Toghill (1968*a*), but most geologists visiting the locality were still guided by the remarkably detailed geological map published by Lapworth in 1878. Following several years' critical work using aerial and ground photographic overlays and modern structural synthesis, Ingham (1979) published a totally revised geological map of Dob's Linn. During the course of this research, Ingham found that several of Toghill's measured sections through the Upper Hartfell and lowermost Birkhill Shales were disrupted structurally and measurements were consequently revised. Ingham's recognition of unbroken sections and several new graptolitic bands in the Upper Hartfell Shale (upper *Complanatus* Band, *Anceps* Band A and the *Extraordinarius* Band), permitted critical faunal recollection of the Moffat Shale. This task was begun by the present author in 1978, leading to a series of taxonomic and biostratigraphical papers covering the top 8 m of the Lower Hartfell Shale (Williams 1982*a*), the *Complanatus* Bands (Williams & Ingham, in prep.), the *Anceps* Bands (Williams 1982), the *Extraordinarius* Band and the basal 2 m of Birkhill Shale (Williams 1983).

These papers confirmed Lapworth's original faith in graptolites as a critical biostratigraphical tool and gave more precise definitions of the zonal boundaries. Of particular importance is the revised, unambiguous definition of the boundary between the *G. persculptus* and *P. acuminatus* Zones, the horizon now defined as the Ordovician–Silurian boundary.

Regional setting, stratigraphy and depositional environment

The geology of the Southern Uplands is dominated by a thick package of monotonous, sparsely graptolitic greywackes, belonging to the Gala Greywacke Group. This is of unequivocal turbidite origin. The underlying Moffat Shale Group is exposed as a series of elongate, narrow, east-west inliers which Lapworth (1878) and Peach & Horne (1899) considered to represent tight, isoclinal anticlines. It is now considered (Webb 1983) that these structures were formed through progressive shearing of early folds. The appearance of simple, reverse faulting postulated by Craig & Walton (1959), Leggett *et al.* (1979), Eales (1979) and other recent workers is due to almost complete removal of the shorter, south-eastern limbs.

An overall younging and progressive lateral change in lithologies from north to south was recognized in the Southern Uplands by Peach & Horne (1899). They divided the regions into three tracts, namely the Northern, Central and Southern Belts. The rock types and age ranges of strata characterizing each belt have since been summarized in detail by Leggett *et al.* (1979), who considered division into ten discrete sequences to be more appropriate. In the most northerly sequences red cherts, siliceous mudstones and pillow basalts of Arenig to Llandeilo age are overlain by Llandeilo–Caradoc greywackes. This succession passes southwards to Llandeilo–Llandovery cherts and black shales overlain by Llandovery greywackes. The diachronous base of the greywackes youngs progressively to the south, with consequently extended black shale deposition. The most southerly sequences of the Southern Uplands are composed entirely of Wenlock greywackes.

Both the structural pattern of the Moffat Shale outcrops and the diachronous base of the greywackes were explained in a model proposed by Mitchell & McKerrow (1975) and expanded by McKerrow *et al.* (1977) and Leggett *et al.* (1979). These authors considered the Southern Uplands to have formed as an accretionary prism over a northerly dipping subduction zone on the northern margin of the Iapetus Ocean. The prehnite-pumpellyite metamorphic facies could have resulted from burial and tectonic processes during such accretion (Oliver *et al.* 1984). Geophysical studies (Powell 1971; Hall *et al.* 1983), however, indicate crystalline, continental material underlying the area, rather than the oceanic basement required for this model. Bluck (1984) discussed this apparently contradictory evidence; he concluded that the Southern Uplands are probably allochthonous. More recently, Needham & Knipe (1986) reiterated the accretionary prism model, but this was considered inadequate by Murphy & Hutton (1986), who concluded that subduction at both Iapetus margins was complete by late Ordovician times and that the Silurian turbidites were deposited in a successor basin.

The Moffat Shale Group is divided into four formations: the Glenkiln Shale, Lower Hartfell Shale, Upper Hartfell Shale and Birkhill Shale (Lapworth 1878). The Glenkiln Shale is composed of an unknown thickness of pale grey and black, heavily silicified argillites. At Dob's Linn the formation is poorly exposed as a series of disconnected, fault-bounded slivers. It is generally unfossiliferous and due to heavy shattering of the competent, siliceous component, even black lithologies rarely yield identifiable graptolites. Useful comparative sections are exposed at the type section of Glenkiln Burn and at several other inliers in the Moffat area (Lapworth 1878; Peach & Horne 1899).

The Glenkiln Shale apparently passes gradationally into the almost continuously black Lower Hartfell Shale, which yields a more abundant graptolite fauna and is over 20 m thick. The lower half of the formation remains highly siliceous; the proportion of chert to black shale decreases upwards throughout the unit, black shale becoming predominant in the upper 5 m.

The overlying Upper Hartfell Shale is composed mostly of monotonous, non-graptolitic, pale grey/green shales and mudstones 28 m thick (Figs 1, 2). Its lower boundary is marked by a transitional 3 cm interval of alternating pale grey and black laminae. Three groups of graptolitic, black shale bands occur within the formation, named the *Complanatus*, *Anceps* and *Extraordinarius* Bands (Ingham 1974, 1979) after their diagnostic zonal assemblages (Fig. 1). Other atypical lithologies include nodular limestones and one detrital limestone. The latter horizon is a very pale grey, coarse-grained limestone 6·5 cm thick, lying 1·5 m below the lower *Complanatus* Band in the banks of the Linn Branch stream. Unfortunately it has been totally recrystallized and affected by strain-induced pressure solution, but it was presumably of detrital origin.

One medium grey nodular limestone, 4 cm thick and lying 2 m above the base of the Upper Hartfell on the North Cliff section, displays uncompacted bioturbation with horizontal to subvertical simple burrows 1–2 mm in diameter. Other nodular horizons present in the Linn Branch section include that known to yield a blind, dalmanitid trilobite 0·1 m below the *Extraordinarius* Band (Ingham 1979) and a second, apparently unfossiliferous bed 0·25 m below the base of the Birkhill Shale. Three of these four limestones were not known prior to recent recollecting for conodont samples (Barnes & Williams, this volume) and other similar horizons in the Upper Hartfell Shale probably still await discovery.

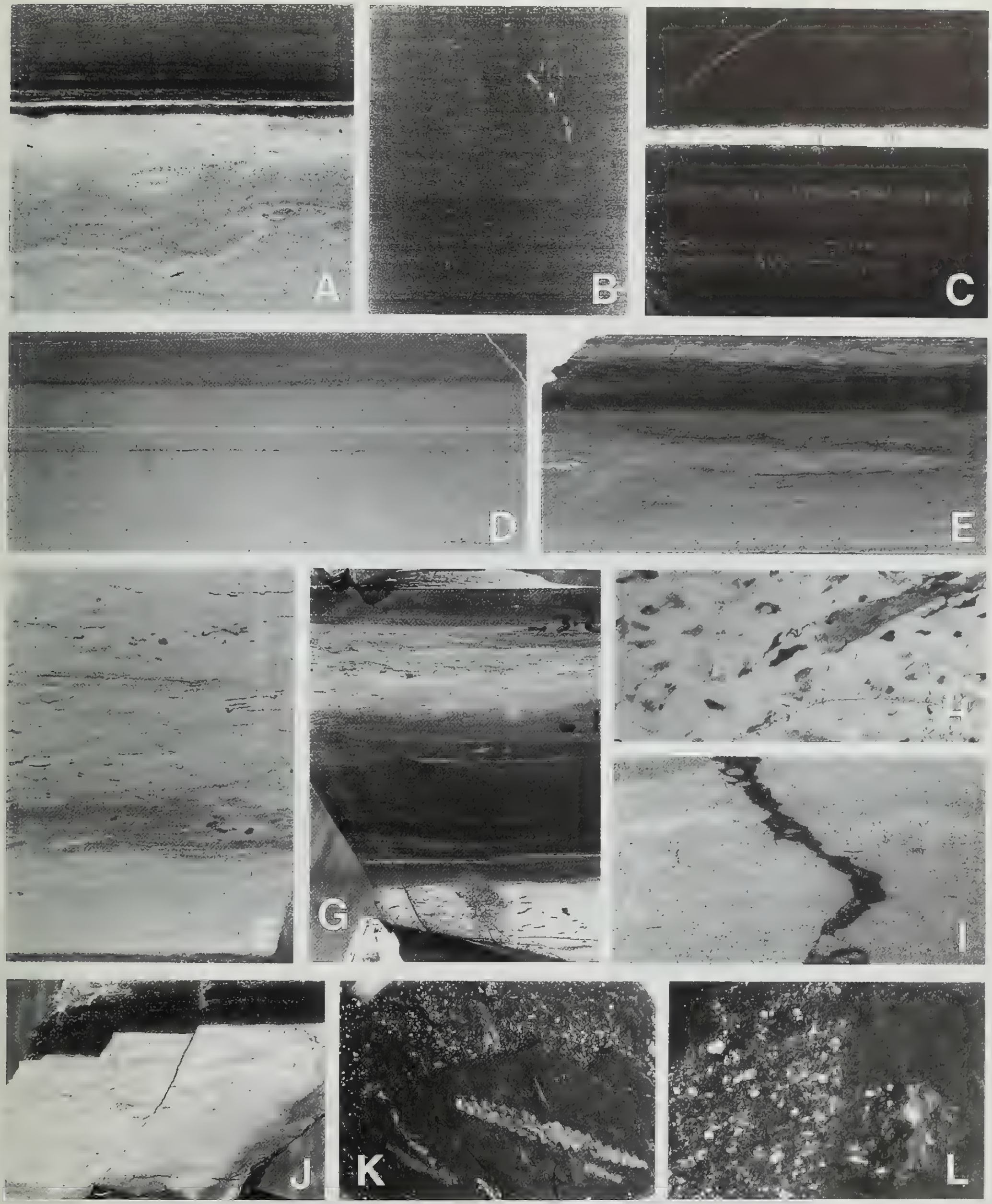

Fig. 2 Sectioned slabs and bedding surfaces from the Moffat Shale at Dob's Linn (all × 2). A. Lower boundary of *Anceps* Band C. B. Typical uniformly laminated black shale, lower Birkhill Shale. C. Black shale, thin metabentonite and micaceous horizons from *Anceps* Band D (note grading shown by metabentonite). D. Uniform pale grey Upper Hartfell Shale lithology, *Anceps* Bands. E. Irregular laminae and compacted bioturbation, base of lower *Complanatus* Band. F. Bioturbation above upper *Complanatus* Band. G. Irregular laminae, compacted bioturbation and low-angle synsedimentary faulting from reversal in Birkhill Shale 0·5 m above base. H. Bedding plane section from base of slab shown in Fig. 2G. I. Black shale injection through pale mudstone, *Anceps* Bands. J. High-angle, post-compactional microfaulting, *Anceps* Band D. K, L. Bedding surfaces with coarse mica flakes, *Anceps* Band D.

The 43 m of Birkhill Shale is composed of black, continuously graptolitic shale in the lower part, with the exception of a temporary reversal to an 'Upper Hartfell' type lithology 0·46–0·56 m above its base. The shales become progressively siltier, less fissile and paler towards the top of the formation, culminating in a transition to coarse turbidites of the overlying Gala Greywacke Group.

The precise depositional environment of the Moffat Shale Group is still uncertain. Lapworth (1897) envisaged black shale formation in a partially restricted 'Sargasso Sea' setting. Later Walton (1963) considered the Moffat Shale to have been deposited on a regional high in a deep ocean environment, explaining the lack of turbidites which are found elsewhere as lateral equivalents and overlying the group.

With the recognition of the Lower Palaeozoic Iapetus Ocean in recent years, it has become evident that the Moffat Shale was deposited within a wide, open ocean of complex history. This suffered continued narrowing throughout the Upper Ordovician and Silurian due apparently to subduction on both northern and southern margins (Moseley 1978, Bluck 1984). The significance of sedimentary features such as postulated winnowing of graptolities, lithological colour alternation, soft-sediment deformation and presence of limited bioturbation (Fig. 2) was discussed by Williams & Rickards (1984). Further observations have emphasized the variation in contacts between pale and black lithologies, from sharp and laminar (Fig. 2A) to gradational and irregular (Figs 2E–G). They have also confirmed the presence of coarse, silty laminae with biotite flakes up to 1 mm diameter, particularly within the *Anceps* Bands of the Upper Hartfell Shale (Figs 2K–L). These strongly suggest a hemipelagic, distal turbidite origin for the sediments, in contradiction to Dewey (1971), Leggett (1980) and Leggett *et al.* (1979), who considered the shale to be of oceanic, truly pelagic origin formed during periods of high eustatic sea level stands. Several black shale sequences elsewhere are known to have been deposited as distal turbidites, including beds of the Burgess Shale of British Columbia (Piper 1972) and of the Cow Head Group, western Newfoundland (Coniglio 1985). It was therefore evident that during Lower Palaeozoic times black shales could form within an unrestricted oceanic setting lacking any degree of restriction, unlike those deposited during Mesozoic and Recent times (e.g. Jenkyns 1978; Stow & Piper 1984).

The presence of metabentonites throughout much of the Moffat Shale indicates sporadic acidic volcanism. Most of these are only laminae or thin beds (Fig. 2C), but they occasionally reach over 5 cm thick. Their lateral impersistence was noted by Williams & Rickards (1984), who suggested variable deposition due to a gently undulating sea floor. It seems likely that the metabentonites were transported by a turbidite mechanism in a similar fashion to the remaining lithologies; they would not, therefore, have significance in terms of proximity to volcanic activity. The single coarse-grained limestone below the lower *Complanatus* Band was probably also deposited by a powerful, carbonate-rich, turbidite flow. Such carbonate detritus was probably derived from a northerly source, such as the sites of fore-arc, shelf and slope deposition at Girvan (Bluck 1984).

No critical sedimentological studies have been carried out on the Moffat Shale at Dob's Linn. With recent advances in both understanding of depositional mechanisms in deep-water, hemipelagic sedimentation (Stow & Piper 1984; Coniglio 1985) and development of new techniques to assist the study of fine-grained sediments, a detailed review of argillites at Dob's Linn and at comparitive sections is now warranted.

Late Ordovician and Early Silurian graptolite biostratigraphy

The following account is based on detailed logging through a trench constructed on the north valley side of the Linn Branch (Figs 3, 4), excepting that of the Lower Hartfell Shale (from the North Cliff trench, Fig. 3) and lower part of the Upper Hartfell Shale, including the *Complanatus* Bands (Linn Branch stream bed).

The uppermost 5 m of the continuously black Lower Hartfell Shale is encompassed within the *Pleurograptus linearis* Zone (Williams 1982*a*). Following this level, 9 m of unfossiliferous grey shale and mudstone belonging to the Upper Hartfell Shale is present before the black,

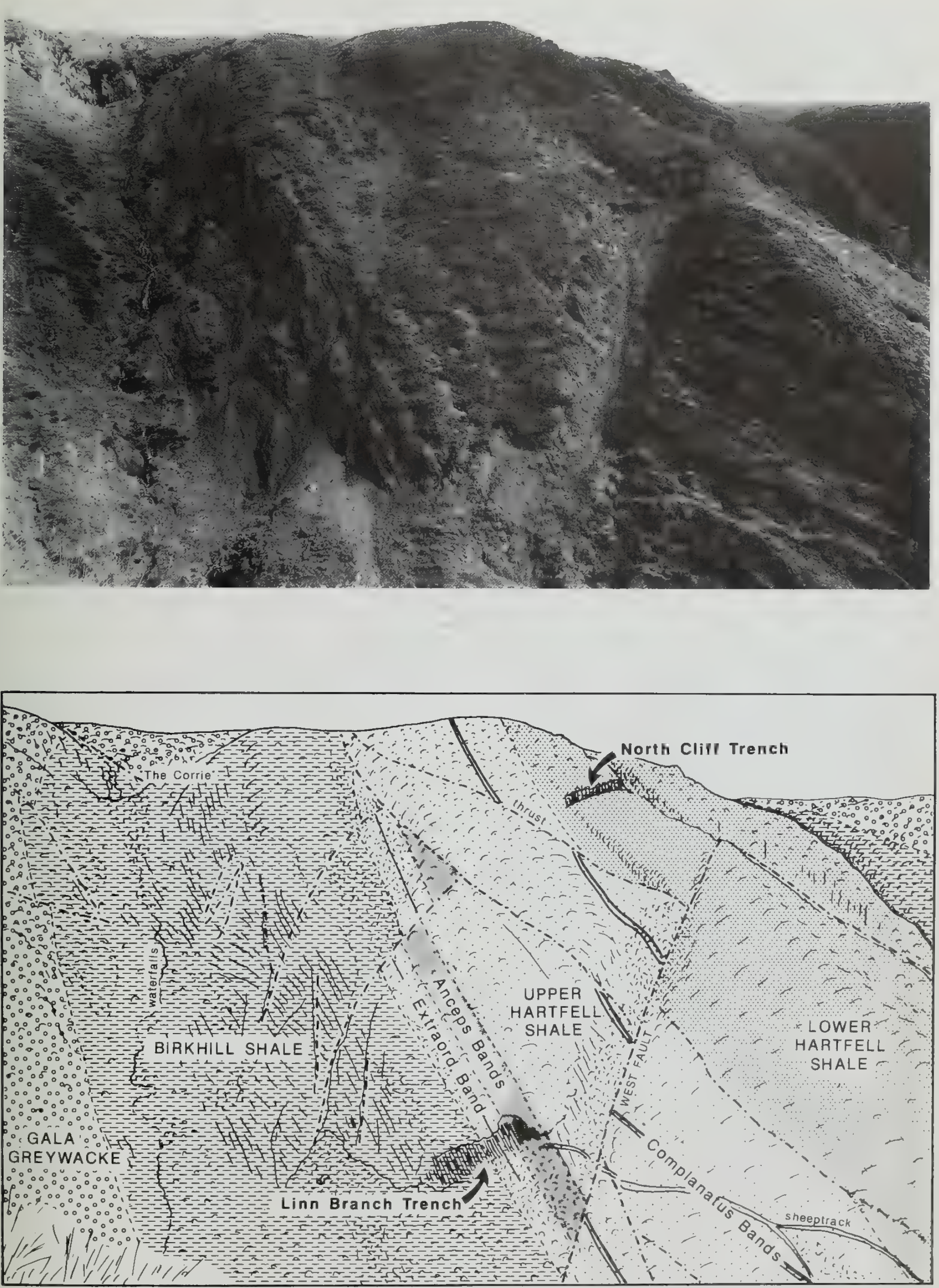
North Cliff Trench
'The Corrie'
thrust
waterfalls
UPPER HARTFELL SHALE
Anceps Bands
Extraord. Band
BIRKHILL SHALE
WEST FAULT
LOWER HARTFELL SHALE
GALA GREYWACKE
Complanatus Bands
Linn Branch Trench
sheeptrack

Fig. 3 Photograph showing northern side of Linn Branch gorge, indicating key collecting localities. Interpretation of geology and structure adapted after Ingham (1974: fig. 25).

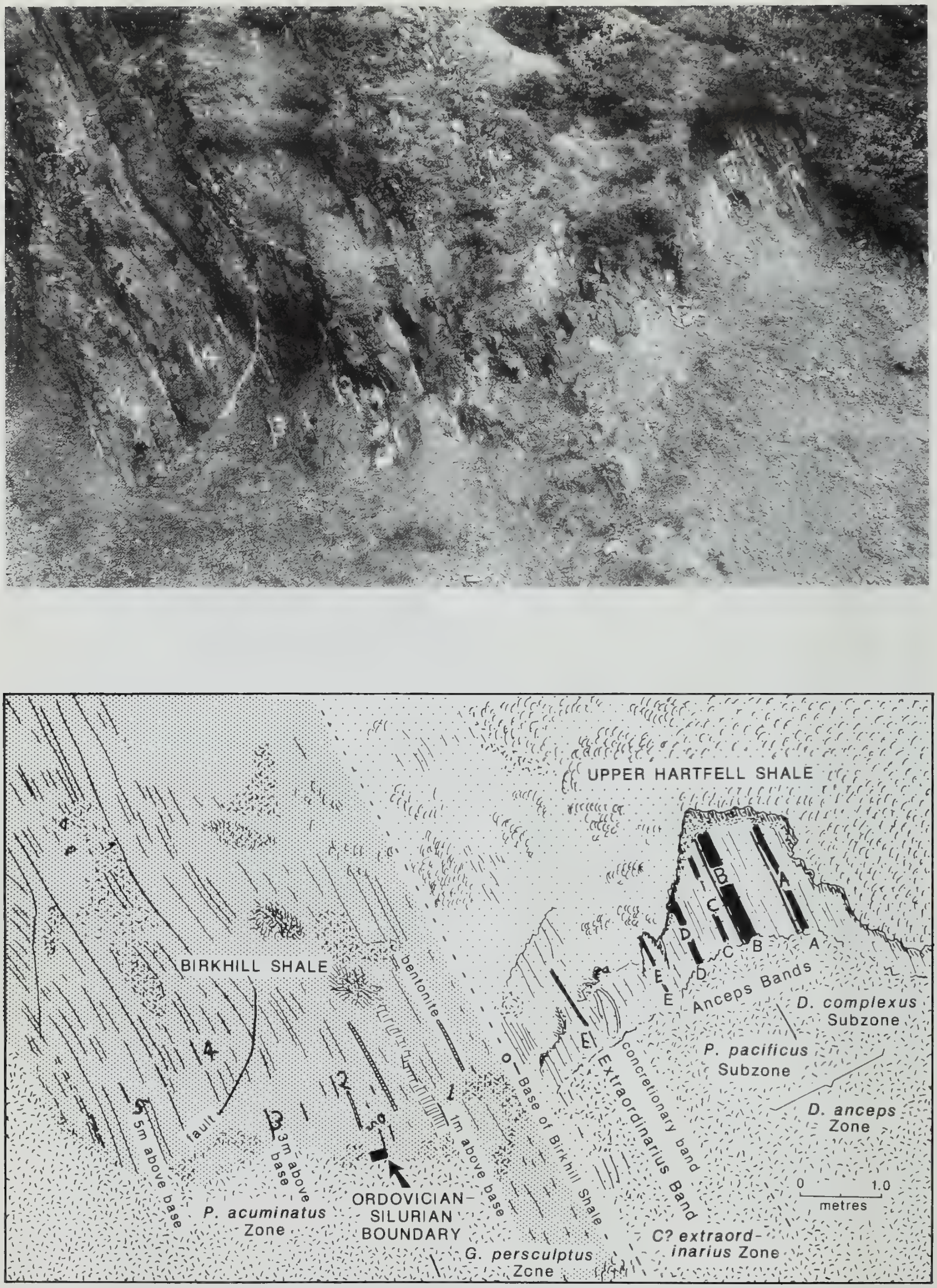
UPPER HARTFELL SHALE
BIRKHILL SHALE
bentonite
B
A
C
D
B
C
A
D
E
E
Anceps Bands
D. complexus Subzone
P. pacificus Subzone
D. anceps Zone
E
Extraordinarius Band
concretionary band
0 — Base of Birkhill Shale
1m above base
1
2
3
4
5
fault
3m above base
5m above base
ORDOVICIAN–SILURIAN BOUNDARY
P. acuminatus Zone
G. persculptus Zone
C? extraordinarius Zone
0
1.0
metres

Fig. 4 Photograph of Linn Branch trench with interpretation, photographed from same position as Fig. 2. Notebook lies on position of Ordovician–Silurian boundary.

graptolitic *Complanatus* Bands are reached. Lapworth (1878: 316; fig. 28) originally recorded '*Dicellograptus forchhammeri, Climacograptus scalaris?* and *Diplograptus truncatus*'. These specimens were subsequently recognized as new taxa, described by Lapworth (1880) and Elles & Wood (1901–18) as *Dicellograptus complanatus, Climacograptus scalaris miserabilis* and *Orthograptus truncatus socialis.* Davies (1929: 18) relocated the graptolite horizon, as did Toghill (1970). Ingham (1974) proved the existence of a second narrow, black seam about 0·4 m above the 4 cm thick, previously recorded band. Williams (1987) records *D. complanatus* Lapworth, *D. minor* Toghill, *C. miserabilis* Elles & Wood, *C. tubuliferus* Lapworth and *O. socialis* (Lapworth) from the lower band. The upper band yields *D. complanatus* and rare specimens of *Orthoretiograptus pulcherrimus* (Keble & Harris). Williams & Lockley (1983) described well preserved specimens of an inarticulate brachiopod, both from within and directly above the upper band, which were assigned to a new genus and species *Barbatulella lacunosa.* Rare, usually fragmented specimens of this brachiopod also occur at several grey mudstone horizons within the following *Anceps* Bands.

The *Anceps* Bands are separated from the *Complanatus* Bands by 13 m of grey barren shale and mudstone. They comprise a series of alternating black and grey shales with common metabentonites, covering an interval which ranges in thickness from 1·6 m on the Main Cliff, to 2·0 m in the Linn Branch trench and 4·5 m in the Long Burn section. The last of these localities is separated from the former two by the Main Fault, and may have been deposited at some distance apart. Other lateral variation in thickness was probably due to deposition on an irregular sea floor and synsedimentary erosion as discussed by Williams & Rickards (1984).

Lapworth (1878: 253, 317) erected the *Dicellograptus anceps* Zone in his major publication on the Moffat Shale, owing to the distinctive nature of the faunal assemblage in the black *Anceps* Bands. Toghill (1970: 6; fig. 1) recorded four black shales; Ingham (1974) however established the presence of five bands or groups of bands, now referred to Bands A to E. The rich, diverse fauna contained within these black shales (Fig. 5) allowed Williams (1982) to divide the zone into the *Dicellograptus complexus* and *Paraorthograptus pacificus* Subzones. In addition to those species' ranges shown on the range chart, rare specimens of *Climacograptus hastatus* Hall and *Glyptograptus posterus* Koren & Tsai have been found in the *D. complexus* and *P. pacificus* Subzones respectively. These taxa confirm correlation with the Australian and Chinese graptolite zonal schemes.

Ingham (1979) was first to discover the *Extraordinarius* Band 0·96 m above *Anceps* Band E. This narrow, dark brown shale contains a sparse graptolite assemblage, identified by Rickards (1979) and Williams (1983) as *Climacograptus? extraordinarius* (Sobolevskaya), *Climacograptus* sp. indet. and *Glyptograptus?* sp. indet. The grey strata separating the *Extraordinarius* Band from *Anceps* Band E is unfossiliferous, with the exception of a nodular limestone 0·1 m below the *Extraordinarius* Bands which yields rare fragmentary specimens of a blind dalmanitid trilobite (Ingham 1979).

The lower boundary of the Birkhill Shale lies 1·17 m above the *Extraordinarius* Band. Following a basal, unfossiliferous black shale interval 0·15 m thick, an abundant but poorly diverse graptolite fauna is present, including *Climacograptus normalis* Lapworth, *C. miserabilis* Elles & Wood and *Glyptograptus? 'venustus* cf. *venustus*' (Legrand). A temporary reversal to alternating grey/green and black shales occurs at 0·46 to 0·56 m above the base. This is followed by black shales yielding a better preserved, more diverse assemblage with the addition of *Glyptograptus* cf. *persculptus* (Salter) and *Glyptograptus? avitus* Davies. Lapworth (1878) referred the basal Birkhill Shale to the *P. acuminatus* Zone. The *G. persculptus* Zone was first separated as a biostratigraphical unit underlying the *P. acuminatus* Zone in central Wales by Jones (1909, 1921), where he also considered it to be lithologically different. Davies (1929) ratified the presence of two distinct zones and recognized the interval equivalent to the *G. persculptus* Zone in both northern England and southern Scotland. It appears that he referred three 'horizons' below the first occurrence of *Parakidograptus acuminatus* (Nicholson) and *Akidograptus ascensus* Davies to the *G. persculptus* Zone at Dob's Linn (1929: 22; fig. 32), but this is not stated unequivocally in the text. Adoption of the *G. persculptus* Zone as a formally defined, distinct biostratigraphic unit at Dob's Linn was not realized prior to Toghill's revision

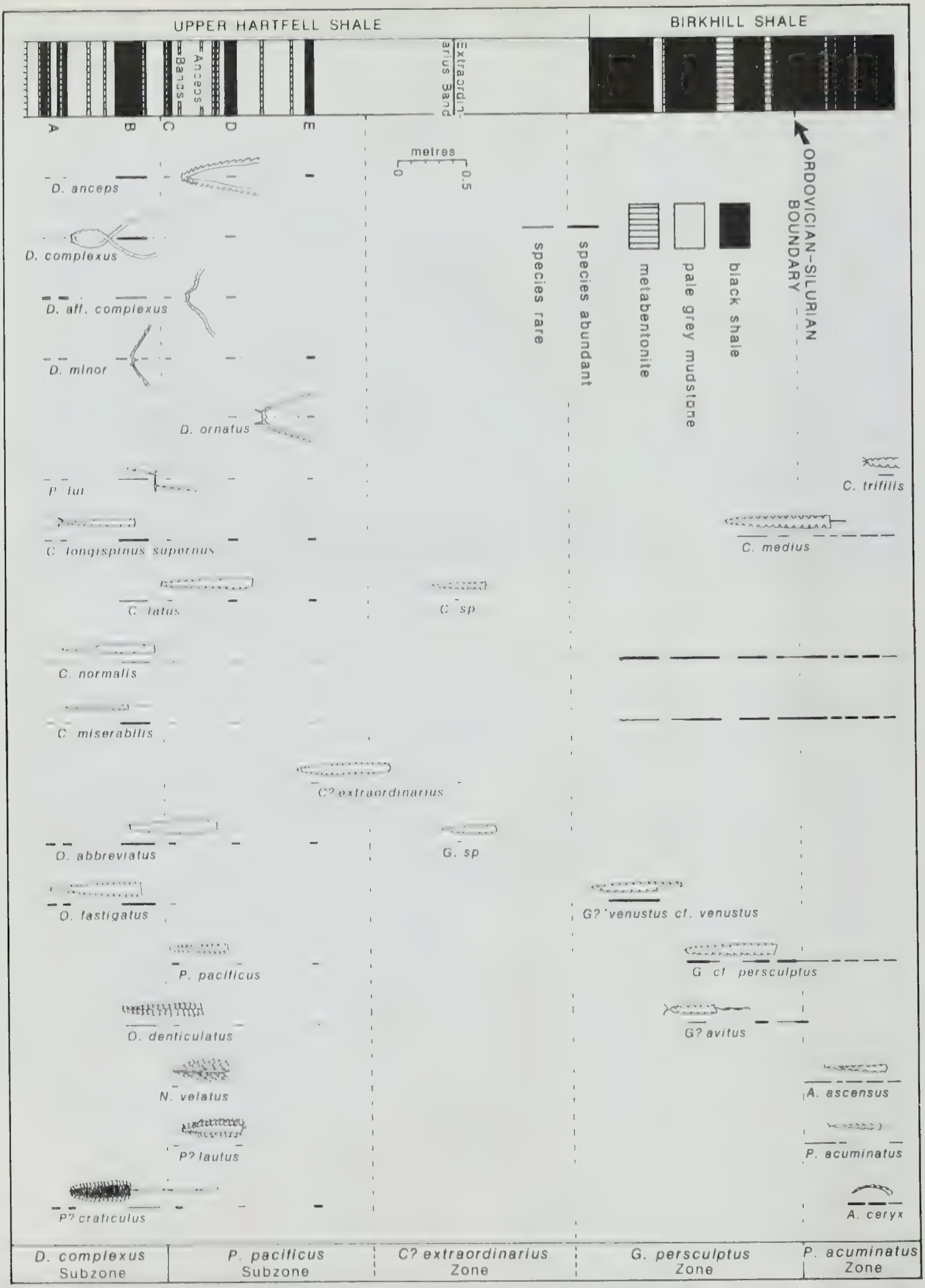

Fig. 5 Detail of sediments and graptolite ranges for the top Upper Hartfell Shale and basal Birkhill Shale.

of the Birkhill Shale in 1968. Rickards (1970) and Hutt (1974) also used this zone as the basal biostratigraphic division of the Lower Silurian Skelgill Formation in northern England.

The base of the *P. acuminatus* Zone is marked by the first appearance of *Akidograptus ascensus* Davies and *Paraorthograptus acuminatus* (Nicholson) *s.l.* at 1·6 m above the base of the Birkhill Shale (Fig. 5). It is this level which has now been adopted as the defined Ordovician Silurian boundary (Cocks 1985), marked by the first occurrence of *A. ascensus*. Most previous publications (e.g. Toghill 1968*a*; Cocks *et al.* 1970) have taken the base of the Birkhill Shale as marking the Ordovician–Silurian boundary. This interval covers a change from grey to black shale, is unfossiliferous and clearly unsuited as a zonal boundary, let alone for an international system boundary stratotype. Similar barren intervals seem to occur at this level in every other graptolitic succession in the world; they are probably related to eustatic sea level changes induced by late Ordovician glaciation in the southern hemisphere (see Rong 1984).

In addition to the problem of barren intervals, faunal changes accompanying the transition between the *G. persculptus* and *C.? extraordinarius* Zones are poorly understood. Few graptolite taxa are present, following the mass extinction at the *D. anceps–C.? extraordinarius* zonal boundary. Elles (1922, 1925) referred the basal interval of the Birkhill Shale to the 'Zone of *Glyptograptus persculptus* and *Cephalograptus acuminatus*'. In the earlier of these publications (1922: 195) she suggested that this lowest Llandovery zone should perhaps be assigned to the Ordovician owing to the lack of monograptids. It is interesting to note that this proposal has now been partially adopted.

Atavograptus ceryx (Rickards & Hutt) occurs 1·9 to 2·3 m above the base of the Birkhill Shale. Recent recollecting indicates that it is probably restricted to such a level low in the *P. acuminatus* Zone at Dob's Linn. *A. ceryx* was first recorded from strata referred to the *G. persculptus* Zone in the English Lake District (Rickards & Hutt 1970), but was later found in the basal *P. acuminatus* Zone of that area (Hutt 1975), in association with *A. ascensus*.

Monograptus cyphus praematurus Toghill and *Atavograptus atavus* (Jones) are the next monograptids found at Dob's Linn, marking the boundary between the *P. acuminatus* and *Cystograptus vesiculosus* Zones (Toghill 1968*a*). Lapworth (1882: 624) recorded an assemblage of '*Climacograptus scalaris*, *Dimorphograptus acuminatus* and *?Monograptus tenuis*' from a section through the Lower Silurian at Girvan, south-west Scotland. Jones (1921: 155) remarked that such an assemblage seemed anomalous for the *P. acuminatus* Zone; it may, however, prove that the monograptid was *A. ceryx* and that the interval was equivalent to the early *P. acuminatus* Zone of Dob's Linn and northern England. Relocation and recollection of Lapworth's horizon could clearly prove significant as a comparative basal Silurian section.

Future research

Dob's Linn has now been adopted as Ordovician–Silurian boundary stratotype, the boundary being set at the base of the *P. acuminatus* Zone 1·6 m above the base of the Birkhill Shale in the Linn Branch trench (Figs 4, 5). This renders necessary ratification and expansion of Williams' (1983) study of the interval. Other outstanding research still required includes:

1. Detailed study of the basal Birkhill Shale at remaining sections of Dob's Linn, and at other comparative localities in the Central Belt of the Southern Uplands.
2. Biostratigraphical and taxonomic revision of the Glenkiln Shale, the lower part of the Lower Hartfell Shale and remainder of the Birkhill Shale, employing continuous, bed-by-bed collecting techniques.
3. Critical sedimentological logging and study of the Moffat Shale at Dob's Linn, with subsequent integration of faunal data, in order to provide a clearer understanding of original depositional setting.

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Conodonts from the Ordovician–Silurian Boundary Stratotype, Dob's Linn, Scotland

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Synopsis

About one hundred poorly preserved conodonts have been collected from surfaces of shale from seven graptolite zones of the Dob's Linn boundary stratotype section, mainly from the *D. anceps* Zone. Attempts to recover conodonts by dissolving siltstones and cherts from the section were unsuccessful. When preserved, the conodont phosphatic material provides Colour Alteration Index values of CAI 5–7, indicating burial temperatures in excess of 300°C. The sparse, low diversity faunas assist in correlating conodont and graptolite zones. *Amorphognathus* sp. and *Scabbardella* sp. cf. *S. altipes* were found in the *G. persculptus* Zone, suggesting that the conodont turnover must lie at least high within this zone. Lowest Silurian strata yielded rare, undiagnostic coniform taxa and an element referred tentatively to *Oulodus*? *kentuckyensis*. The results encourage further efforts in retrieving conodonts from graptolitic shale sequences, but the precise correlation of the conodont turnover with respect to the defined base of the Silurian remains in question.

Introduction

The Ordovician–Silurian boundary was finally designated in 1985 at 1·6 m above the base of the Birkhill Shale in the Linn Branch section of Dob's Linn, southern Scotland, at the base of the *Parakidograptus acuminatus* Zone (Williams 1983 and this volume; Cocks 1985). Detailed work on the rich graptolite faunas has been carried out by a number of previous researchers, especially Lapworth, Elles & Wood, Toghill and Williams (see Williams 1983, this volume). The section, however, has yielded no other biostratigraphically useful fossils in abundance; there are rare inarticulate brachiopods (Williams & Lockley 1983) and a species of a blind dalmanitid trilobite. Lamont & Lindström (1957) reported conodonts from cherts in the Southern Uplands of Scotland, including Dob's Linn, but only gave identifications and details of the Arenig and Llandeilo faunas.

One critical problem in the debate concerning the definition of the Ordovician–Silurian boundary and subsequent selection of a stratotype was that few candidate sections contained both graptolites and conodonts. At the level of the *G. persculptus* and *P. acuminatus* Zones (Fig. 1) in particular, there are difficulties in correlating the graptolite and conodont zones and the two respective extinction events (e.g. Barnes & Bergström, this volume). It is, therefore, both encouraging and important to report in this paper the discovery of conodonts at several levels in the Dob's Linn boundary stratotype section.

While scanning shale surfaces under the microscope during the investigation of graptolites, Williams observed a number of microfossils which have since been identified by Barnes. Further collections were made by Williams in 1985; this time, in addition to the scanning of shale surfaces, samples of shales, siltstones and cherts were processed through a variety of standard chemical rock digestion techniques employed for conodonts (e.g. acetic and hydrofluoric acids; bleach). The latter results were disappointing in that most lithologies appeared to be barren of conodonts, although this may have been due to inadequate preservation (see below). The remaining new collections revealed many additional conodont horizons, but yielded few diagnostic elements from new stratigraphical levels. This project however demonstrates that conodonts are present, and moderately abundant at some horizons, in graptolitic shales deposited in a deep oceanic environment which has been interpreted as an accretionary prism (McKerrow *et al.* 1979; see other recent interpretations by Needham & Knipe 1986 and

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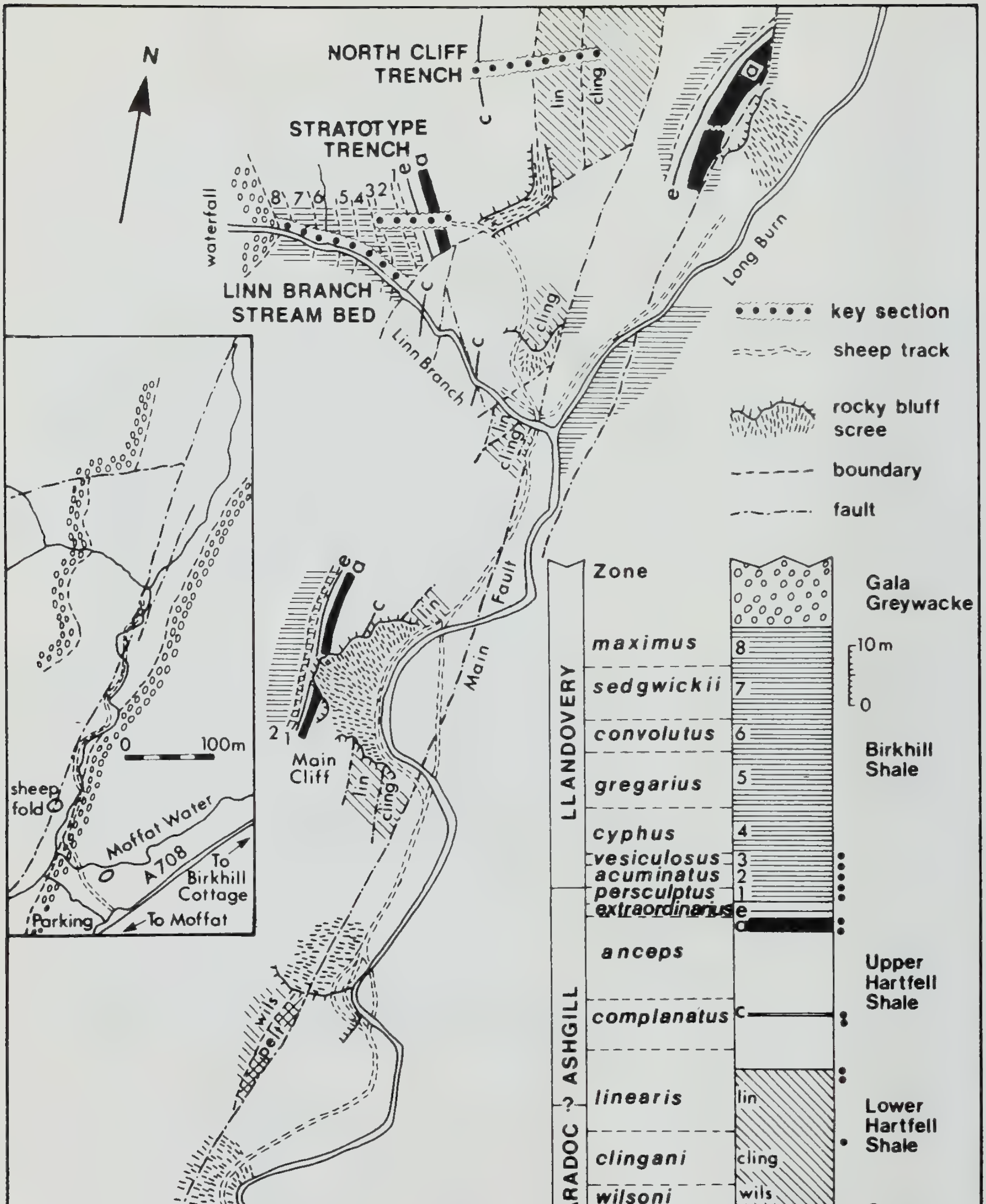
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gregarius
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acuminatus
persculptus
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Birkhill Shale
Upper Hartfell Shale
Lower Hartfell Shale
Glenkiln Shale
• conodonts
e Extraordinarius Band
a Anceps Bands
c Complanatus Bands

Fig. 1 Simplified geological map and stratigraphical section of Dob's Linn, showing position of conodont localities and horizons (after Williams 1980).

Murphy & Hutton 1986). Careful microscopical examination of similar shales in other sequences should reveal many new conodont faunas and assist integration of graptolite and conodont biostratigraphic zonation schemes.

Results

Following the discovery of the microfossils, re-examination of earlier material, together with the new shale collections, has involved the study of several hundred surfaces for conodonts. Conodonts and rare scolecodonts are present. The conodonts always occur as isolated elements; no fused clusters or natural assemblages were discovered. The elements are poorly preserved, typically being fractured by tectonic stretching and commonly with only part of the phosphatic skeletal material preserved. This may, in part, explain the difficulty in obtaining identifiable conodonts from dissolved samples. For some, only an external mould remains, but latex casts have been successfully made which permit specific identifications (e.g. Pl. 1, fig. 10; Pl. 2, fig. 12). The conodonts provide Colour Alteration Index values of CAI 5–7. This is in agreement with the general high thermal values reported elsewhere in the Southern Uplands of Scotland by Bergström (1980), indicating burial temperatures exceeding 300°C.

About one hundred conodont elements have been recognized, the majority of which are identifiable only to generic level. The diversity of the fauna is low, but zonal species are present. Nearly all the conodonts come from Ordovician strata, in particular the *D. anceps* Zone; unfortunately, conodonts are especially rare near the Ordovician–Silurian boundary.

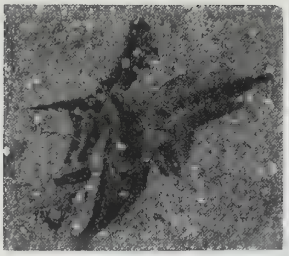
Hartfell Shale conodonts

Most of the conodonts from the Dob's Linn section come from the Hartfell Shale. They were recovered at various levels within the *Dicranograptus clingani*, *Pleurograptus linearis*, *Dicellograptus complanatus* and *D. anceps* Zones, but principally from the latter zone. Details of stratigraphy, together with a revision of the graptolite faunas from the *D. clingani*, *P. linearis*, and *D. anceps* Zones, have been published by Williams (1982*a*, *b*). Conodonts from the *D. clingani* Zone were not identifiable; those from the *P. linearis* Zone included two specimens of *Amorphognathus superbus* (Rhodes) from 1·1–1·2 m and 0·3–0·45 m below the top of the Lower Hartfell Shale, several specimens of *Amorphognathus* sp. and a specimen of *Walliserodus* unidentifiable to species level.

The precise level at which *A. superbus* evolved into *A. ordovicicus* (i.e. base of the *A. ordovicicus* Zone) in terms of graptolite zones remains to be established. This zonal boundary appears to lie within the upper Pusgillian Stage or lower Cautleyan Stage (Bergström 1971, 1983; Orchard 1890; Bergström & Orchard 1985), although Savage & Bassett (1985) tentatively suggest a late Caradoc age. In North America, this boundary occurs in the lower Maysvillian Stage (Sweet & Bergström 1971). The *D. clingani–P. linearis* zonal boundary is approximately equivalent to, or slightly predates, the base of the earliest Ashgill Pusgillian Stage (Williams & Bruton 1983). The samples yielding *A. superbus* are from the top of the Lower Hartfell Shale (mid *P. linearis* Zone; Williams 1982*a*: fig. 3) which probably falls within the Pusgillian Stage.

A single identifiable conodont was recovered from the *D. complanatus* Zone of the Upper Hartfell Shale, namely *Amorphognathus ordovicicus* Branson & Mehl from the lower Complanatus Band. At Myoch Bay in the Girvan area, southern Scotland, the *D. complanatus* Zone of the Upper Whitehouse Group also yields shelly fossils of Pusgillian age (Ingham 1978; Harper 1979). Conodonts from these strata (Sweet & Bergström 1976: 135–136; Bergström & Orchard 1980) do not allow a zonal assignment. It must be emphasized that the material at hand comprises only a single, poorly preserved amorphognathodontiform element; this limited evidence suggests that the *A. ordovicicus* Zone boundary lies within the Pusgillian rather than the Cautleyan.

The *D. anceps* Zone is recognized in the Upper Hartfell Shale by a series of thin black shales assigned to Anceps Bands A–E (e.g. Williams 1982*b*). These contain the most abundant conodont fauna from the Dob's Linn section. No significant difference was observed in the conodont fauna of the various bands except in terms of relative abundance. Band A yielded rare speci-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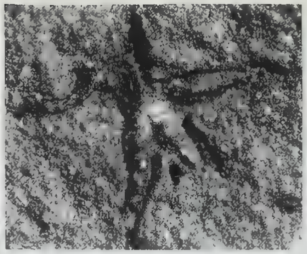
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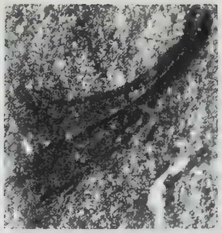
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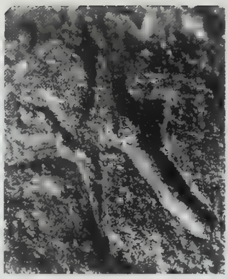
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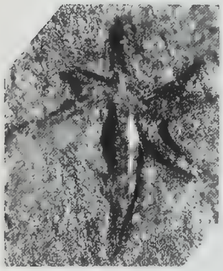
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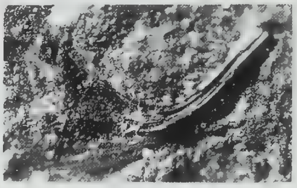
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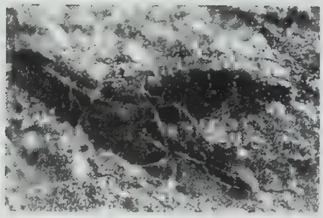
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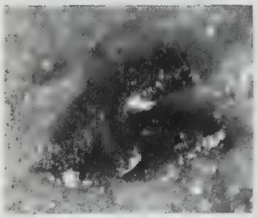
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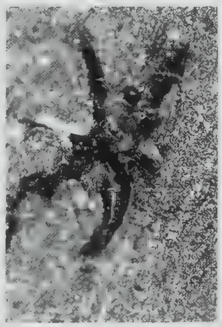
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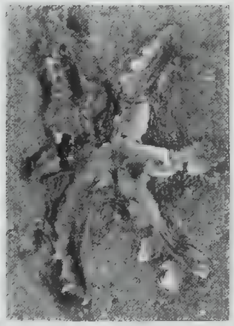
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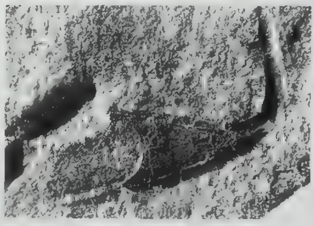
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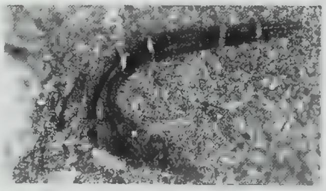
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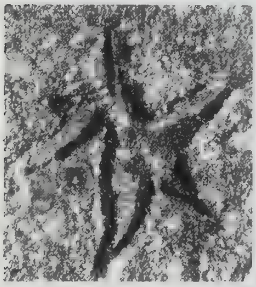
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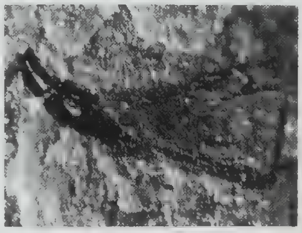
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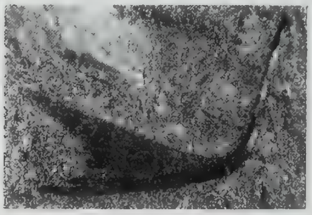
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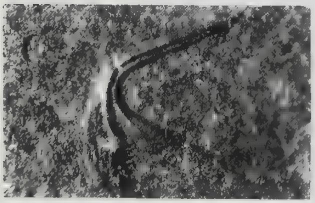
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mens assignable to only two species: *Amorphognathus ordovicicus* and *Protopanderodus liripipus* Kennedy, Barnes & Uyeno. Band B produced conodonts referred to *A.* sp. cf. *A. ordovicicus*, *Scabbardella altipes* (Henningsmoen) and an oistodontiform element that probably belonged to *Hamarodus europaeus* (Serpagli). Band C contained only *P. liripipus*, and Band D yielded *A.* sp. cf. *A. ordovicicus* and *S. altipes*; both had only rare fragmentary conodonts. Band E contained slightly more specimens including *Amorphognathus* sp., *P. liripipus* and *S. altipes*. The *D. anceps* Zone therefore yields conodonts belonging to the *A. ordovicicus* Zone. Orchard (1980) recovered *H. europaeus* from only Rawthyan and Hirnantian strata although the range of this species has now been extended into the Cautleyan by Barnes & Bergström (this volume).

No conodonts were recovered from the 1-cm black shale *Extraordinarius* Band of the top Upper Hartfell Shale, which yields *C.*? *extraordinarius* Zone graptolites of probable mid-Hirnantian age (Williams 1983).

Birkhill Shale conodonts

The Birkhill Shale includes the upper part of the top Ordovician *G. persculptus* Zone, the basal Silurian *Parakidograptus acuminatus* Zone and subsequent Llandovery graptolite zones (Toghill 1968; Williams 1983). The lower few metres of the Birkhill Shale is the critical interval from which turnover conodonts need to be recovered, but unfortunately no especially diagnostic taxa were found.

In strata of the *G. persculptus* Zone, the few specimens observed were all coniform except for one slightly crushed and distorted specimen of *Amorphognathus* sp. from 0·12–0·2 m above the base of the Birkhill Shale. The coniform taxa include *Dapsilodus obliquicostatus* (Branson & Mehl) and *Scabbardella* sp. cf. *S. altipes*. The latter occurs at 0·95 m above the base of the Birkhill Shale. Neither *Amorphognathus* nor *Scabbardella* are known with certainty from Silurian strata. This limited evidence, based on rare, poorly preserved specimens, suggests that most of the *G. persculptus* Zone may lie below the main Ordovician–Silurian conodont turnover (see discussion by Barnes & Bergström, this volume).

The *P. acuminatus* Zone, beginning at 1·6 m above the base of the Birkhill Shale, and the overlying *Cystograptus vesiculosus* Zone contained a few conodonts assigned to *Dapsilodus obliquicostatus* and *Decoriconus* sp. In addition a single, small, poorly preserved ligonodiniform element was found at 1·75 m above the base of the Birkhill Shale. The form of the lateral

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PLATE 1 Conodonts from the Lower and Upper Hartfell Shale, Dob's Linn, Scotland.

Figs 1, 2 *Amorphognathus superbus* (Rhodes) × 35. 1, dextral blade element, upper view. HM Y155. 1·1 m below top of Lower Hartfell Shale, *P. linearis* Zone. North Cliff. 2, dextral blade element, upper view of mould. HM Y157. 0·3–0·45 m below top of Lower Hartfell Shale, North Cliff.

Fig. 3 *Walliserodus* sp. × 70. Lateral view. HM Y201. Top of Lower Hartfell Shale, North Cliff.

Figs 4, 5, 13 *Amorphognathus ordovicicus* Branson & Mehl. × 35. Upper Hartfell Shale. 4, sinistral blade element, upper view of mould. HM Y159. Lower *Complanatus* Band. 5, dextral blade element, upper view of mould. HM Y107. *Anceps* Band A. Long Burn. 13, dextral blade element, upper view of mould. HM Y129. *Anceps* Band D. Main Cliff.

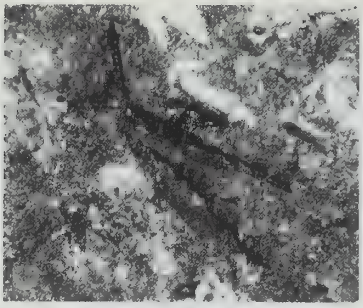
Figs 6, 12, 16 *Protopanderodus liripipus* Kennedy, Barnes & Uyeno. × 55. Upper Hartfell Shale. 6, scandodontiform element. HM Y109a. *Anceps* Band A, Long Burn. 12, symmetrical element. HM Y121. *Anceps* Band C. Main Cliff. 16, scolopodontiform element. HM Y135. *Anceps* Band E. Linn Branch.

Figs 7, 11, 14, 15, 18 *Scabbardella altipes* Henningsmoen. Lateral views. × 55. Upper Hartfell Shale. 7, ?acodontiform element. HM Y203. *Anceps* Band B. Linn Branch. 11, distacodontiform element. HM Y112. *Anceps* Band B. Main Cliff. 14, acodontiform element. HM Y202. *Anceps* Band D. Linn Branch. 15, distacodontiform element. HM Y126. *Anceps* Band D. Long Burn. 18, distacodontiform element. HM Y204. 40 cm above *Anceps* Band E, Linn Branch.

Fig. 8 *Hamerodus europaeus* (Serpagli). × 55. Oistodontiform element. HM Y205. *Anceps* Band B. Linn Branch.

Figs 9, 10 *Amorphognathus* sp. cf. *A. ordovicicus* Branson & Mehl. × 35. 9, dextral blade. Upper view of mould. HM Y114b. *Anceps* Band B. Main Cliff. 10, latex cast of HM Y114b (Fig. 9).

Fig. 17 *Amorphognathus* sp. × 35. Dextral blade element, upper view of mould. HM Y136. *Anceps* Band E. Linn Branch.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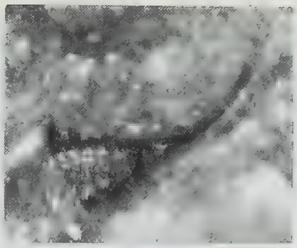
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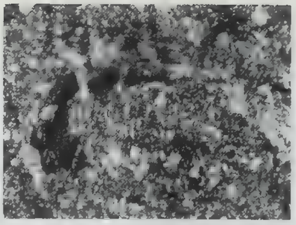
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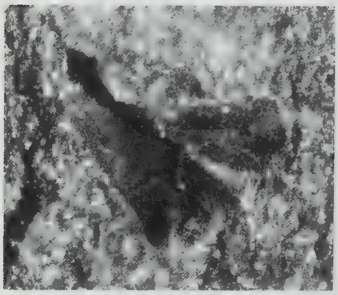
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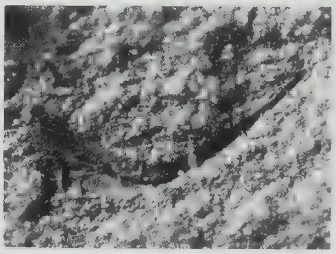
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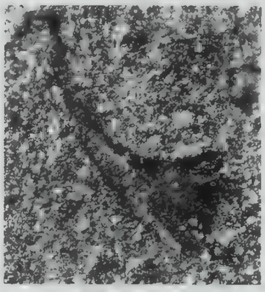
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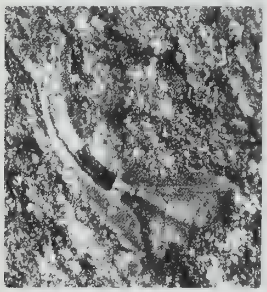
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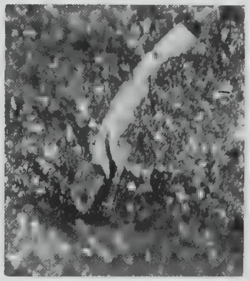
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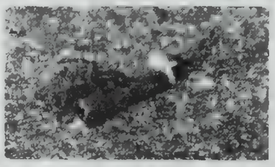
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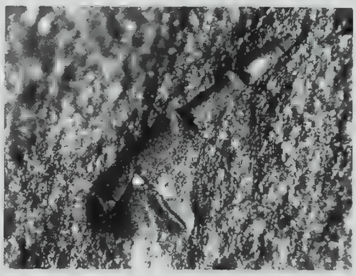
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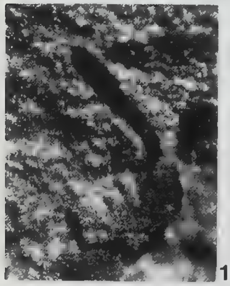
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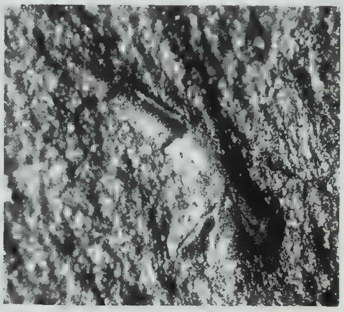
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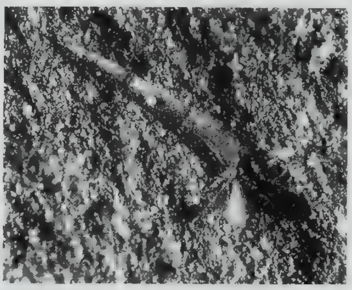
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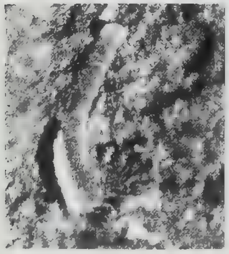
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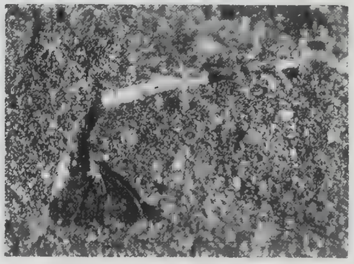
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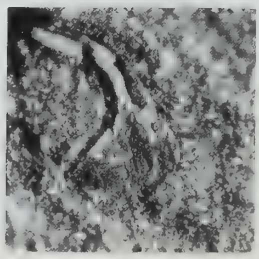
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process extends into the shale but its shape is revealed by a latex cast. The element is assigned tentatively to *Oulodus? kentuckyensis* (Branson & Mehl). The latter species is known only from Silurian strata elsewhere (e.g. Anticosti Island, McCracken & Barnes 1981).

Summary

About 100 conodont elements have been observed on shale surfaces from the Dob's Linn boundary stratotype section. Most are from black shales, but occasional specimens also occur within paler grey shales and siltstones. The elements are poorly preserved, fractured and commonly occur as moulds; the Colour Alteration Index values are in the range of CAI 5–7 indicating burial temperatures exceeding 300°C. Identification of most elements can be made only to generic level; a selection of the better specimens are here illustrated (Figs 2, 3) but the photography for many proved difficult and not all details of micromorphology could be reproduced. The diversity of the faunas is low, typically 3–5 species per graptolite zone interval. This may be expected in the deep oceanic environment of the Hartfell Shale and Birkhill Shale, but is probably also related to the limited material discovered. Siltstone, shale and chert samples were also processed chemically but yielded no identifiable conodonts. Although the sparse fauna and poor preservation must be taken into account, the following biostratigraphic conclusions may be drawn from this study.

1. *Amorphognathus superbus* is present in the *Pleurograptus linearis* Zone near the top of the Lower Hartfell Shale (based only on amorphognathodontiform, not holodontiform elements). *Amorphognathus ordovicicus* occurs in the *Dicellograptus complanatus* Zone of the Upper Hartfell Shale. This suggests that the *A. superbus–A. ordovicicus* zonal boundary is not far removed from that of the *P. linearis* and *D. complanatus* Zones and lies within the Pusgillian Stage.

2. Most of the conodonts come from the *Dicellograptus anceps* Zone; all the *Anceps* Bands A–E of the Upper Hartfell Shale yielded specimens, which are indicative of the *A. ordovicicus* Zone. Conodonts also occur at several grey, silty, non-graptolitic horizons during this interval.

3. No conodonts were recovered from the 1-cm black shale of the *Climacograptus? extraordinarius* Zone.

4. The lower 1·6 m of the Birkhill Shale, belonging to the *Glyptograptus persculptus* Zone, contained two poor specimens of *Amorphognathus* sp. and *Scabbardella* sp. cf. *S. altipes*, known only from Ordovician strata. This suggests that the major conodont turnover (Barnes & Bergström, this volume) occurred at a level equivalent to at least high in the *G. persculptus* Zone.

PLATE 2 Conodonts from the Birkhill Shale, Dob's Linn, Scotland. Figs 1–16 arranged in order of stratigraphical occurrence of specimens. *G. persculptus* Zone (Figs 1–9); *P. acuminatus* Zone (Figs 10–14); *C. vesiculosus* Zone (Figs 15, 16).

Fig. 1 *Amorphognathus* sp. × 35. Upper view, distorted specimen. HM Y142. 0·12–0·2 m above base of Birkhill Shale.

Figs 2, 3, 5, 9 *Dapsilodus* sp. Lateral views. 2, HM Y206. × 90. 0·55 m above base of Birkhill Shale. 3, HM Y207. × 55. 0·95 m above base of Birkhill Shale. 5, HM Y208. × 80. 0·95 m above base of Birkhill Shale. 9, HM Y209. × 55. 1·5 m above base of Birkhill Shale.

Figs 4, 6 *Scabbardella altipes* Henningsmoen. Lateral views. × 55. 4, HM Y210. 0·95 m above base of Birkhill Shale. 6, HM Y211. 0·95 m above base of Birkhill Shale.

Figs 7, 15, 16 *Dapsilodus obliquicostatus* (Branson & Mehl). Lateral views. 7, HM Y213. × 70. 1 m above base of Birkhill Shale. 15, HM Y214. × 55. 5 m above base of Birkhill Shale. 16, HM Y215. × 55. 5·5 m above base of Birkhill Shale.

Fig. 8 *Drepanoistodus* sp. × 70. Lateral view. Drepanodontiform element. HM Y212. 1 m above base of Birkhill Shale.

Figs 10, 12, 13 *Decoriconus* sp. × 55. Lateral views. 10, HM Y216. 1·75 m above base of Birkhill Shale. 12, HM Y217. 1·75 m above base of Birkhill Shale. 13, latex cast of HM Y217 (Fig. 12).

Figs 11, 14 cf. *Oulodus? kentuckyensis* (Branson & Branson). × 105. Lateral views. 11, ligonodiniform element. HM Y218. 1·75 m above base of Birkhill Shale. 14, latex cast of HM Y218 (Fig. 11).

5. Silurian conodonts from the *Parakidograptus acuminatus* and *Cystograptus vesiculosus* Zones include mostly coniform taxa (*Dapsilodus*, *Decoriconus*) which cross the systemic boundary at other localities. A poor single element assigned tentatively to *Oulodus*? *kentuckyensis*, which elsewhere is known only from Silurian strata, was found in the *P. acuminatus* Zone.

These results suggest that more attention should be made to recover conodonts from shales, particularly in graptolitic shale sequences. The above data must be used with caution until more material is discovered. However, the situation is perhaps analogous to the presence of poorly preserved, rare graptolites within the conodont-rich Anticosti Island carbonate boundary sequence (McCracken & Barnes 1981; Riva, this volume). It remains one of the future challenges to find a boundary sequence that yields both well preserved and abundant, biostratigraphically diagnostic conodonts and graptolites across the systemic boundary.

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Preliminary acritarch and chitinozoan distributions across the Ordovician–Silurian boundary stratotype at Dob's Linn, Scotland

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Synopsis

Palynomorph distribution has been investigated across the Ordovician–Silurian boundary at Dob's Linn, where the Hartfell Shale and Birkhill Shale are well exposed. Samples were taken from the *anceps*, *extraordinarius*, *persculptus*, and *acuminatus* graptolite Biozones at the stratotype Linn Branch section and also the Main Cliff. Graptolite debris is the dominant component of the organic fraction, but acritarchs, chitinozoans and scolecodonts also occur in small numbers. Although it has not been possible to define the position of the Ordovician–Silurian boundary by microflora, the presence of palynomorphs indicates that detailed sampling might provide the stratigraphical resolution necessary to do this.

At Dob's Linn in the Southern Uplands of Scotland, continuous sections through the Hartfell and Birkhill Shales (Caradoc to Llandovery) bracket the Ordovician–Silurian boundary. These shales are replaced vertically by greywackes (the Gala Greywackes) in the *maximus* Zone. Fault bounded tracts showing similar transitions are common in the Southern Uplands. Systematic variations in the regional timing of this transition, and the complex younging relationships between and within tracts, are thought to reflect the progressive growth of an accretionary prism (McKerrow *et al.* 1977) during closure of the Iapetus Ocean. The 90 m of Hartfell and Birkhill Shales exposed here (Williams 1981) represent a substantially condensed sequence, as an equivalent sequence a hundred kilometres to the west, at Girvan, is over 3000 m thick.

This is a preliminary report of the distribution of acritarchs and chitinozoans across the newly formalized Ordovician–Silurian boundary at Dob's Linn. Data from the *anceps*, *extraordinarius*, *persculptus* (all Ordovician) and *acuminatus* (Silurian) graptolite Biozones are presented. Palynomorphs were recovered from hydrofluoric and hydrochloric acid-etched residues and studied using the scanning electron microscope, or transmitted light microscope. Whilst graptolites are common at Dob's Linn, and other fossils, such as scolecodonts, have been found sporadically, this is the first major palynological survey that has been undertaken on the Ordovician and the basal Silurian there.

The older Upper Hartfell Shale is a sequence (28 m thick) of finely bioturbated massive grey mudstones (Williams & Rickards 1984), with subordinate thin black shale bands (two *complanatus* bands, five *anceps* bands and one *extraordinarius* band), and metabentonite horizons. The Birkhill Shale (48 m) comprises a laminated, pyritous, black shale with abundant graptolites, and representing the *persculptus* to *maximus* Zones. The systemic boundary of the Ordovician–Silurian has been fixed at the base of the *acuminatus* graptolite Biozone, 1·6 m above the base of the Birkhill Shale (Cocks 1985).

Samples have been collected from two localities spanning the boundary, the Main Cliff and the Linn Branch section (the world stratotype of the Ordovician–Silurian Boundary). At Main Cliff the *wilsoni* to *acuminatus* graptolite Zones are exposed, and although some strike slip faulting has caused repetition of the upper *anceps* and *extraordinarius* black shale bands, the beds are consistently the right way up (Williams 1980). At the Linn Branch, the *anceps* to *maximus* Zones are present, and although the beds are overturned, the stratigraphy is not complicated by repetition. To date sampling has concentrated on the *extraordinarius* and *anceps* Zones. However, work in progress aims to characterize the distribution of palynomorphs across the boundary.

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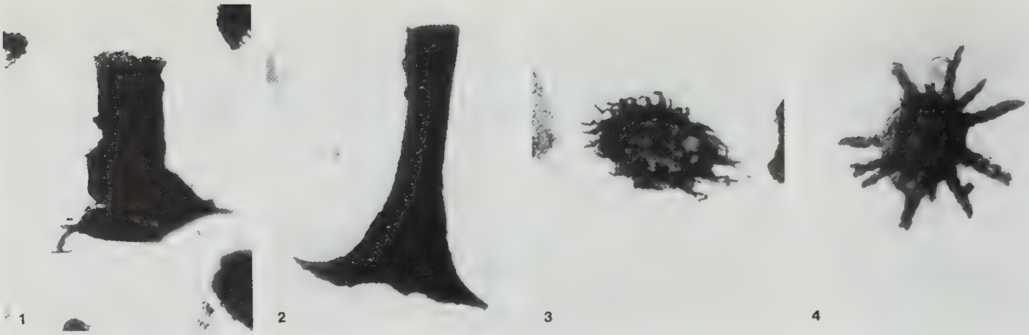


Figs 1–4 Chitinozoans and acritarchs from Dob's Linn. 1, *Ancyrochitina ancyrea* (Eisenack 1931) Eisenack 1955. SU/DL/41, *acumunatus* Zone, Main Cliff, × 250. 2, *Cyathochitina kukersiana* (Eisenack 1934) Eisenack 1965. SU/DL/9, *anceps* Zone, Main Cliff, × 250. 3, *Solisphaeridium nanum* (Deflandre 1945) Turner 1984. SU/DL/12, *anceps* Zone, Main Cliff, × 530. 4, *Diexallophasis* sp. 1. SU/DL/10, *anceps* Zone, Main Cliff, × 470.

Both groups of palynomorphs are unevenly distributed throughout the two sections although they are generally more abundant at Main Cliff. Acritarchs appear to be more important and better preserved in the grey mudstones, while chitinozoans appear to be more common in the black shales, although this is not always the case. Palynomorph colour varies from grey to black within a single sample, and probably reflects differences in wall thickness.

Acritarchs can be divided into several groups (Downie *et al.* 1963): (a) **Sphaeromorphs** which are spherical. These are of limited biostratigraphical use as can be seen in Figs 1 and 2, and will not be mentioned further; (b) **Acanthomorphs** which have spines or processes; (c) **Herkomorphs** which have crested ridges forming polygonal fields; (d) **Polygonomorphs** which have a limited number of processes, usually between three and five; and (e) **Netromorphs** which are generally fusiform in shape. The Dob's Linn samples are noticeably dominated by acanthomorph acritarchs and only a few samples contain representatives of the other groups.

Anceps Zone

Six samples have been studied from Main Cliff (only one of which is a grey mudstone) and sixteen acritarch and chitinozoa taxa have been found (Fig. 5). The chitinozoans *Cyathochitina campanulaeformis* (Eisenack), *C. kukersiana* (Eisenack) and *Rhabdochitina gallica* Taugourdeau all suggest a Caradoc to Ashill age. *Hercochitina* cf. *turnbulli* Jenkins has previously been described from the Caradoc of Oklahoma (Jenkins 1969), but only one poorly preserved specimen was found at Dob's Linn. The acritarch *Solisphaeridium nanum* (Deflandre) Turner ranges from Arenig to Devonian and is therefore a poor biostratigraphical indicator. Of the other acritarchs recovered *Stellechinatum brachyscolum* Turner has been described only from the Caradoc of Shropshire (Turner 1984), and *Veryhachium reductum* (Deunff) Jekhowsky from the Tremadoc to the Silurian. *Diexallophasis* sp. 1 has also been found from the Silurian *sedgwickii* Zone and is probably a new species (pers. comms Molyneux 1986). Thus palynomorphs indicate an Upper Ordovician age for the *anceps* Zone, primarily on the evidence of chitinozoan distribution. Samples from the *anceps* Zone at the Linn Branch section have yielded no palynomorphs and this is attributed to the extreme weathering of this part of the section.

Extraordinarius Zone

The chitinozoans and the acritarch *Veryhachium corpulentum* Colbath found in this zone (Figs 5, 6) suggest a Caradoc to Ashgill age, although the acritarchs *Veryhachium lairdii* and *V. reductum* both range from Lower Ordovician to Silurian in age. The Linn Branch section has only yielded two non-sphaeromorph acritarchs: the acanthomorphs *Baltisphaeridium* sp. 1 and *Armoricanium* sp. 2 (Fig. 6).

GRAPTOLITE ZONE	SAMPLE NUMBER	LITH.	CHITINOZOANS	ACANTHOMORPH ACRITARCHS	SPHAEROMORPH ACRITARCHS	OTHER ACRITARCHS
ACUMINATUS	SU DL 41	≡	ANCYROCHITINA ANCYREA ANCYROCHITINA SP 1 KALOCHITINA SP 1		LEIOSPHAERIDIA SP 1	
PERSCULPTUS	SU DL 40	≡	RHABDOCHITINA MAGNA		L SP 1	DICTYOTIDIUM SP 1
EXTRAORDINARIUS	SU DL 38			MULTIPLICISPHAERIDIUM SP 1	L SP 1 L. SP 2 SYNSPHAERIDIUM SP 1	VERYHACHIUM LAIRDII V CORPULENTUM V REDUCTUM
	SU DL 17	·	CYATHOCHITINA HYMENOPHORA	MICRHYSTRIDIUM SP 1 MICRHYSTRIDIUM SP 2	L. SP 1	V. SP 1 ACTINOTODISSUS SP 1
	SU DL 16	≡			L SP 1	
	SU DL 15			MULTIPLICISPHAERIDIUM SP 1	L SP 1 L SP 2	
ANCEPS	SU DL 14	≡		MULTIPLICISPHAERIDIUM SP 1		
	SU DL 13	≡			L SP 1 L SP 2	
	SU DL 12	·		SOLISPHAERIDIUM NANUM STELLECHINATUM BRACHYSCOLUM MICRHYSTRIDIUM SP 1	L. SP 1 L SP 2	
	SU DL 11	≡			L SP 1 L. SP ?	
	SU DL 10	≡		GONIOSPHAERIDIUM SP 1 DIEXALLOPHASIS SP 1 MULTIPLICISPHAERIDIUM SP 2 MICRHYSTRIDIUM SP 1, M SP 3	L. SP. 1 L SP 2	VERYHACHIUM REDUCTUM AREMORICANIUM SP 1
	SU DL 9	≡	CYATHOCHITINA KUKERSIANA C CAMPANULAEFORMIS RHABDOCHITINA GALLICA HERCOCHITINA CF TURBULLI		L. SP 1 L. SP 2	

Fig. 5 Distribution of acritarchs and chitinozoans at Main Cliff, Dob's Linn. In column 3 (lithology), horizontal lines indicate a black shale sample, and the dots represent a grey mudstone.

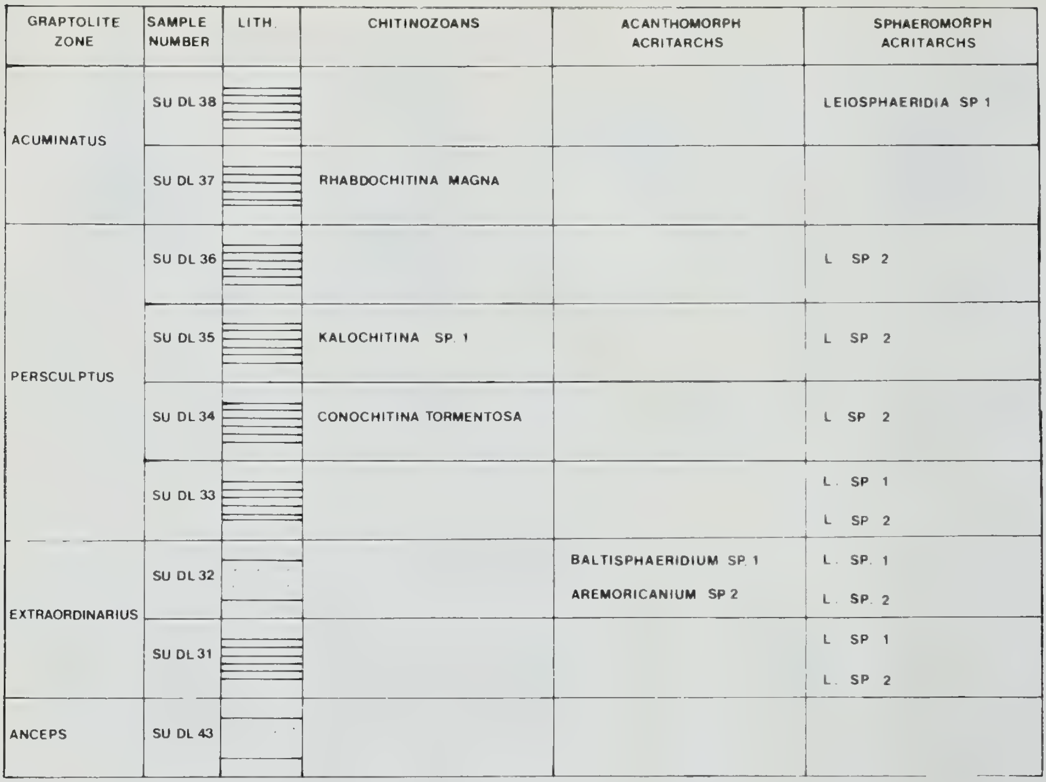
Persculptus Zone

At Main Cliff the chitinozoan *Rhabdochitina magna* Eisenack and the herkomorph acritarch *Dictyotidium* sp. 1 have been found, while two samples from the Linn Branch section have yielded *Kalochitina* sp. 1 and *Conochitina tormentosa* Taugourdeau. This assemblage suggests a Caradoc to Ashgill age, although *Rhabdochitina magna* is known to range into the Llandovery.

Acuminatus Zone

One sample from Main Cliff has yielded 24 specimens of the important Lower Silurian form *Ancyrochitina ancyrea* (Eisenack) Eisenack and a single specimen of *Kalochitina* sp. 1. At the Linn Branch *Rhabdochitina magna* is found, and both this species and *Kalochitina* sp. 1 extend across the boundary, and are thus of little biostratigraphical use as boundary markers.

Because of the long range of most species the distributions of acritarchs and chitinozoans are less refined biostratigraphical indicators than those of graptolites. The sample from the *acuminatus* Zone can be dated accurately as Lower Silurian, while all other samples which yielded an unequivocal age determination are of Upper Ordovician age. It is important to note that the chitinozoans have proved most useful in this survey and that they are often very abundant in the black shales. As the boundary is within the Birkhill Shale, it is possible that bed by bed processing will yield sufficient chitinozoan taxa to determine the position of the Ordovician–Silurian boundary accurately in terms of the microflora. As palynomorphs often occur in rocks which lack datable macrofossils, even a crude biostratigraphical zonation based on chitinozoans would have considerable use in word-wide correlation.



GRAPTOLITE ZONE	SAMPLE NUMBER	LITH.	CHITINOZOANS	ACANTHOMORPH ACRITARCHS	SPHAEROMORPH ACRITARCHS
ACUMINATUS	SU DL 38				LEIOSPHAERIDIA SP 1
	SU DL 37		RHABDOCHITINA MAGNA		
PERSCULPTUS	SU DL 36				L SP 2
	SU DL 35		KALOCHITINA SP. 1		L SP 2
	SU DL 34		CONOCHITINA TORMENTOSA		L SP 2
	SU DL 33				L. SP 1 L SP 2
EXTRAORDINARIUS	SU DL 32			BALTISPHAERIDIUM SP. 1 AREMORICANIUM SP 2	L. SP. 1 L. SP. 2
	SU DL 31				L SP 1 L. SP 2
ANCEPS	SU DL 43				

Fig. 6 Distribution of acritarchs and chitinozoa at the Linn Branch section, Dob's Linn. Lithology symbols as in Fig. 5.

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Ordovician–Silurian junctions in the Girvan district, S.W. Scotland

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Synopsis

The Ordovician–Silurian boundary at Girvan is represented by a variety of unconformable contacts; the basal Silurian rocks both overstep and overlap the upper Ordovician strata south and southwestwards. The most complete section across the junction is in a regressive shelly facies located north of the Girvan valley in the Craighead inlier. The Hirnantian High Mains Formation contains a moderately diverse *Hirnantia* fauna within channel fill sandstones. The overlying basal Silurian unit, the middle Rhuddanian Mulloch Hill Conglomerate, was deposited in submarine canyons at a variety of depths and contains an entrained *Cryptothyrella* fauna. The continuing regression evident across the junction and the facies patterns in the lowermost Silurian are related to the local emergence of fault-bounded blocks.

Introduction

The Ordovician and Silurian rocks of the Girvan district, S.W. Scotland contain a wide variety of siliciclastic sediments, together with locally diverse shelly and graptolite faunas; deposition occurred in a proximal fore-arc environment (Bluck 1983). In contrast to the graptolite facies of the Ordovician–Silurian boundary sections in the shale inliers of the Southern Uplands, the most stratigraphically complete junction section at Girvan is in a shelly facies.

Lapworth's detailed study of the Girvan succession (1882) was largely confirmed by the similarly substantial researches of Peach & Horne (1899). But neither study was aware of the terminal Ordovician unit, the High Mains Formation; thus the marked contrast between the faunas of the Ladyburn Mudstones of the Upper Drummuck Group and those of the Mulloch Hill Group led Lapworth (1882: 622) to consider the apparent hiatus between the top of his Ardmillan Series and the base of his Newlands Series to represent 'the grandest palaeontological break in the entire Girvan succession'.

In a detailed appraisal of the Drummuck Group, Lamont (1935: 294) noted the presence of a hitherto unrecognized unit of buff-weathering sandstone overlying the Drummuck Group and containing a distinctive shelly fauna. He considered the unit, the High Mains Sandstone, to represent the base of the Mulloch Hill Group and moreover (Lamont 1935: 289) suggested a correlation with the lower Llandovery. From this unit he briefly described and figured specimens of his new genus *Hirnantia*, which he based on material of *Orthis sagittifera* M'Coy from both the High Mains Sandstone and the Hirnant beds of Bala, north Wales, and noted the presence of *Meristella* sp. (*Hindella crassa incipiens*). Subsequently, Lamont (1949) described the trilobite *Flexicalymene scotica* from the High Mains Sandstone and modified his views on the correlation of the unit to include the possibility of a Hirnantian age. Ingham & Wright (1970) subsequently emphasized the presence of key elements of the terminal Ordovician *Hirnantia* fauna and concluded a correlation with the Hirnantian Stage.

Harper (1979*b*) noted the presence of two distinct associations of the *Hirnantia* fauna within the High Mains Sandstone and suggested the inapplicability of the term 'community' to contain the marked diversity of associations within the *Hirnantia* fauna. The formation has been described and mapped in detail and bulk samples of the two shelly associations investigated (Harper 1981). The thirteen taxa of brachiopod are currently being described (Harper 1984 and in preparation), whilst Owen (1986) has completed a monographic study of the five taxa of trilobites.

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The junction sections

The basal Silurian strata both overstep and overlap the upper Ordovician rocks of the district south and southwestwards (Cocks & Toghill 1973). The most stratigraphically complete boundary section is thus north of the Girvan valley in the Craighead inlier (Fig. 1) whilst the largest hiatus is developed in the coastal exposures south of the Girvan Valley and southwest of the main outcrop (Fig. 2).

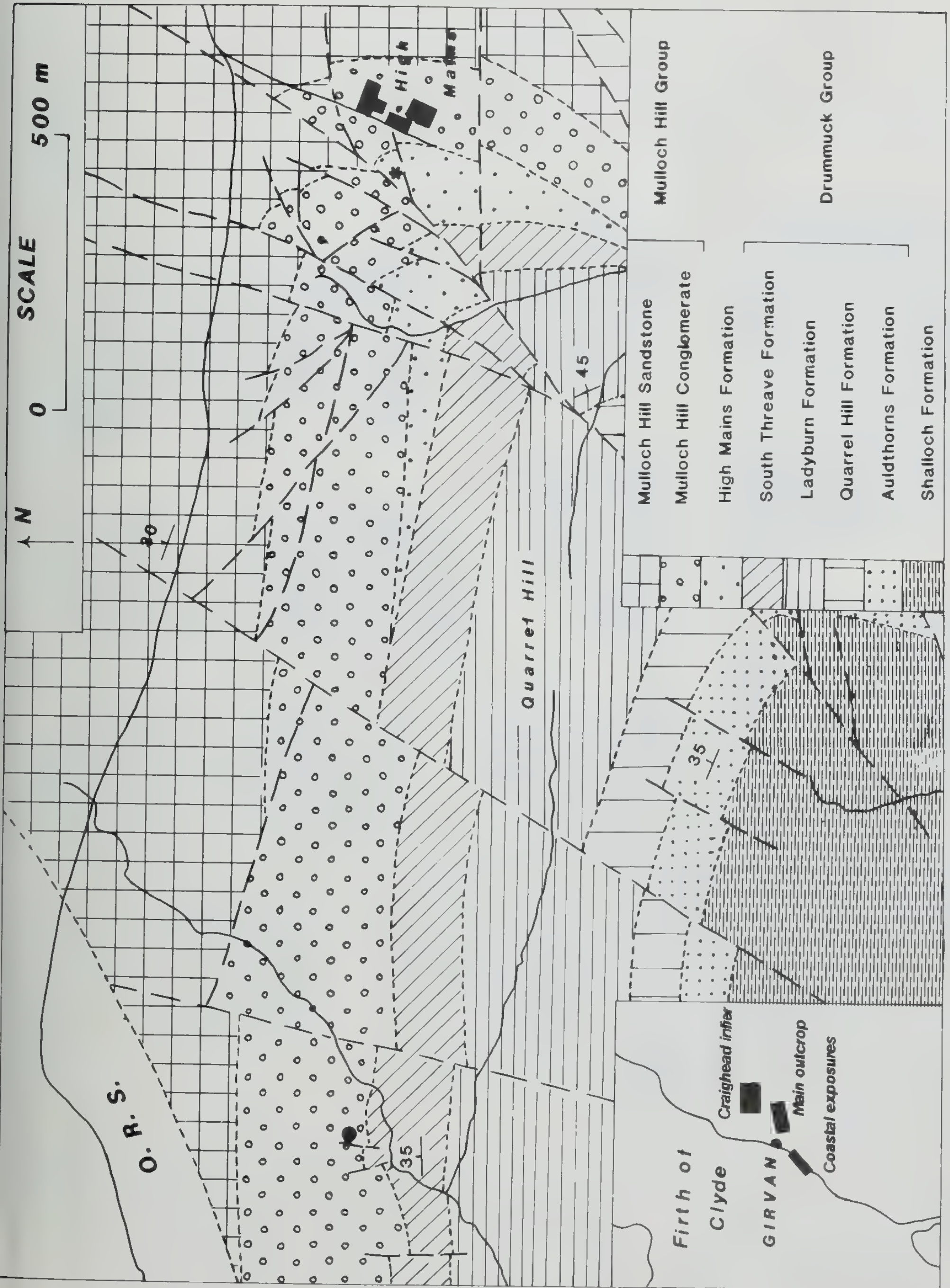
(i) *Craighead inlier.* The terminal Ordovician unit, the High Mains Formation, crops out in the vicinity of High Mains farmhouse (Fig. 1). The unit is poorly exposed, and the detailed outcrop pattern (Harper 1981) was investigated by trenching and mechanical digging. The formation contains two associations of the *Hirnantia* fauna and a Hirnantian age is indicated. The High Mains Formation is overlain by the Mulloch Hill Conglomerate (the Lady Burn Conglomerate of Cocks & Toghill, 1973) but although the junction is not exposed it is assumed to be fairly sharp with a slight angular discordance.

(ii) *Main Outcrop.* The main outcrop of Silurian rocks in the Girvan district extends from Saugh Hill approximately northeast to Straiton (Cocks & Toghill 1973: fig. 1). The presence of major bedding-parallel structures have locally tectonized the shale units and may be responsible for the variation of thicknesses, along strike, of several of the formations. The junction of the Silurian with the underlying Ordovician is exposed on the west bank of Penwhapple Burn (National Grid ref. NX 2327 9769) some 500 metres downstream from Penwhapple Bridge (Cocks & Toghill 1973: fig. 4). Here, the local base of the Silurian is represented by the Tralorg Formation. At the junction the succession is inverted; however, the Tralorg Formation appears to overlie conformably grey micaceous sandstones and shales of the Shalloch Formation; the junction is apparently tectonized as are the shales within the underlying Shalloch Formation. In an adjacent quarry, graptolites of the *anceps* Zone indicate a middle Ashgill age for this part of the Shalloch Formation. Both units dip steeply south.

(iii) *Coastal Exposures.* The two main coastal exposures of the Ordovician–Silurian junction clearly demonstrate the southward overstep and overlap of the basal Silurian units. At the northernmost of the two exposures, the Haven (Cocks & Toghill 1973: fig. 3), the Craigskelly Conglomerate overlies the Shalloch Formation unconformably. However, farther south on Woodland Point the Woodland Formation unconformably overlies lower horizons of the Shalloch Formation, although pockets of Craigskelly Conglomerate lie between the two.

Faunal and facies changes at the Ordovician–Silurian junction

As noted above, the most complete boundary section is near High Mains farmhouse in the central part of the Craighead inlier (Fig. 1). The highest Ordovician strata in the Girvan district, in ascending order the Shalloch Formation, the Drummuck Group and the High Mains Formation, are sporadically exposed and the latter two units are locally highly fossiliferous (Figs 3–22). Within the Drummuck Group a variety of shelly associations dominated by brachiopods have been noted (Harper 1979*b*), and are currently under detailed description, together with the continuing documentation of the brachiopod taxa (Harper 1984 and in preparation). The associations are thought to have inhabited a spectrum of environments upslope and adjacent to the proximal parts of a submarine fan system. The highest strata of the group, the upper Rawtheyan South Threave Formation (Harper 1982), contain highly fossiliferous sandstones (the Ladyburn Starfish Beds of the Farden Member) and probable mudflow units (the Cliff Member); nevertheless background sedimentation is represented by bedded green mudstones and occasional siltstones containing low diversity faunas of minute inarticulate, enteletacean and plectambonitacean brachiopods. The boundary with the overlying Hirnantian High Mains Formation, although not exposed, is assumed to be fairly sharp. The High Mains Formation consists of fine-medium and medium grained quartz sandstones. The beds are massive with an apparent lack of internal sedimentary structures; horizons of shelly debris


SCALE
0
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30
45
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35
Quarrel Hill
High Mains
Mulloch Hill Sandstone
Mulloch Hill Conglomerate
Mulloch Hill Group
High Mains Formation
South Threave Formation
Ladyburn Formation
Quarrel Hill Formation
Auldthorns Formation
Shalloch Formation
Drummuck Group
Firth of Clyde
GIRVAN
Craighead inlier
Main outcrop
Coastal exposures

Fig. 1 Detailed map of the central part of the Craighead inlier; the black asterisk indicates the position of the High Mains trench (fossiliferous localities H1 and H2 of Harper 1981), whilst the black dot marks the site of a fossiliferous sandstone within the Mulloch Hill Conglomerate. The inset, bottom left, shows the approximate positions of three main areas of Silurian outcrop discussed.

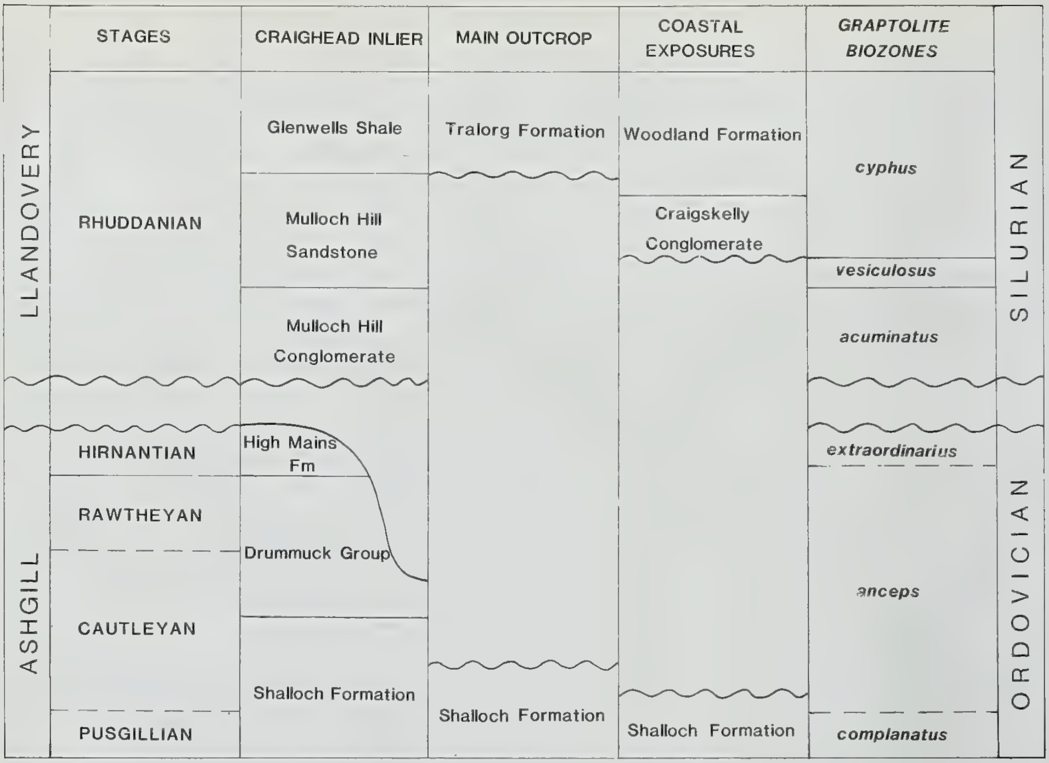
STAGES
CRAIGHEAD INLIER
MAIN OUTCROP
COASTAL EXPOSURES
GRAPTOLITE BIOZONES
LLANDOVERY
RHUDDANIAN
Glenwells Shale
Tralorg Formation
Woodland Formation
cyphus
SILURIAN
Mulloch Hill Sandstone
Craigskelly Conglomerate
vesiculosus
Mulloch Hill Conglomerate
acuminatus
ASHGILL
HIRNANTIAN
High Mains Fm
extraordinarius
RAWTHEYAN
Drummuck Group
CAUTLEYAN
anceps
ORDOVICIAN
Shalloch Formation
Shalloch Formation
Shalloch Formation
PUSGILLIAN
complanatus

Fig. 2 Correlation of Ordovician–Silurian junction sections across the Girvan district with each other and the established shelly stages and graptolite biozones.

are developed at various levels in the formation. The lower 2 m of the formation exposed in the High Mains trench (Harper 1981: 250) comprises grey-green fine-medium grained, thinly bedded sandstones, whilst the upper 5·5 m is a hard medium-grained, thickly-bedded sandstone. Changes in grain size, bedding characteristics and faunal composition indicate a minor regression within the sequence. In view of the incomplete exposure and an apparent absence of

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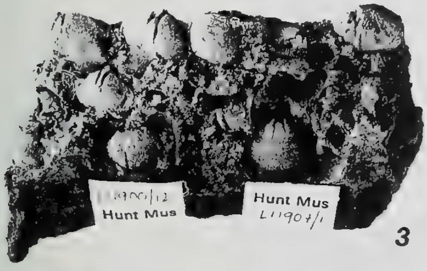
Figs 3–22 Brachiopods and trilobites from the High Mains Formation (Hirnantian), High Mains trench, Girvan. Repository: Hunterian Museum, Glasgow. Fig. 3, fossiliferous block of the High Mains sandstone dominated by internal moulds of both pedicle and brachial valves of *Hindella crassa* (J. de C. Sowerby) *incipiens* (Williams) and crinoid ossicles, × 1. Figs 4, 8, *Plaesiomys* aff. *porcata* (M'Coy). 4, HM L12238, latex cast of internal mould of brachial valve, × 2; 8, HM L12239, latex cast of external mould of pedicle valve, × 2. Figs 5, 9, 13, *Eochonetes* cf. *advena* Reed. 5, HM L12115, internal mould of pedicle valve, × 4; 9, HM L12117, latex cast of internal mould of brachial valve, × 4; 13, HM L12118, latex cast of external mould of pedicle valve, × 3. Figs 6, 7, 10, 11, *Hindella crassa* (J. de C. Sowerby) *incipiens* (Williams). 6, 10, HM L12242, latex cast and internal mould of pedicle valve, both × 2; 7, HM L12244a, external mould of brachial valve, × 3; 11, HM L12244b, latex cast of internal mould of brachial valve, × 3. Figs 12, 14–16, *Eostropheodonta* aff. *hirnantensis* (M'Coy). 12, HM L12105, latex cast of internal mould of brachial valve, × 2; 14, HM L12104, internal mould of pedicle valve, × 1; 15, HM L12103, latex cast of external mould of pedicle valve, × 2; 16, HM L12653, latex cast of internal mould of brachial valve, × 2. Figs 17, 18, *Hemiarges extremus* Owen, HM A16153, external mould and latex cast of cranidium, both × 2. Figs 19–21, *Hirnantia sagittifera* (M'Coy). 19, HM L12654, latex cast of brachial valve exterior, × 2; 20, 21, HM L1986, latex cast and internal mould of brachial valve, both × 2. Fig. 22, *Achatella* cf. *truncatocaudata* (Portlock), HM A16152, internal mould of cephalon, × 2.

sedimentary structures, palaeoenvironmental analysis of the High Mains Formation is equivocal. Nevertheless the thickness, geometry and lithology of the unit are compatible with deposition within channels which developed on the deeper parts of the shelf and the upper parts of the slope. Such environments (Dott & Bird 1979) may be characterized by apparently massive and structureless sandstones comprising channel fills in the order of 25 m thick. Elsewhere, various modes of channelling characterize predominantly argillaceous upper Ashgill sequences; these developed during the time of regression in response to the end Ordovician glacio-eustatic event (Brenchley & Newall 1980). At Girvan, however, a fall in sea level in excess of the 50–100 m estimated (Brenchley & Newall 1980: 34) is required and thus additional tectonic controls must be invoked.



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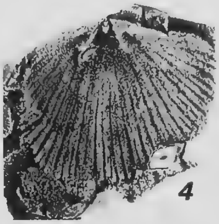
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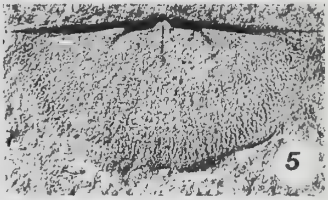
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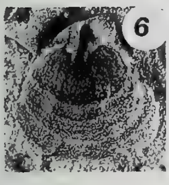
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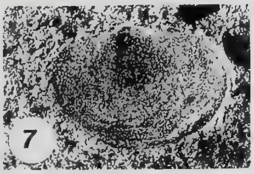
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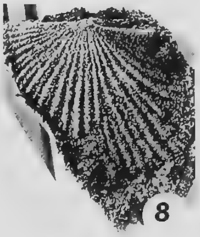
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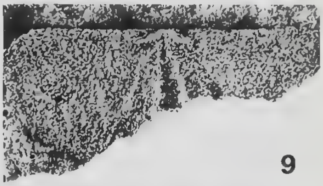
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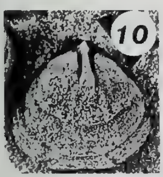
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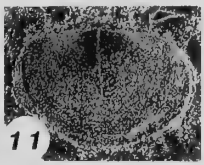
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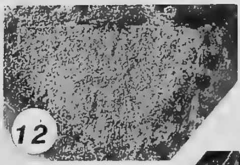
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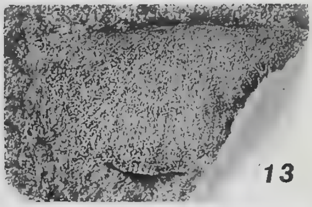
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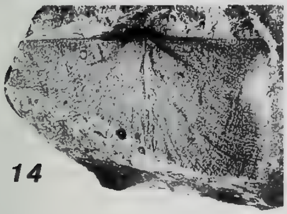
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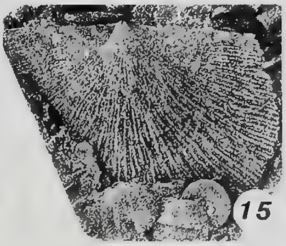
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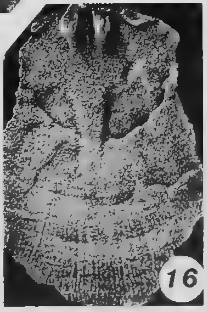
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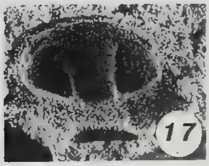
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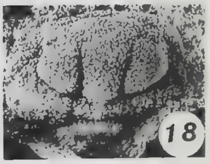
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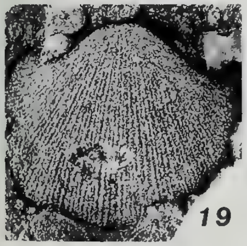
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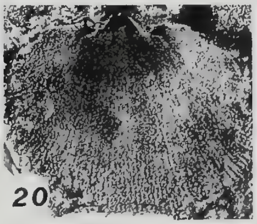
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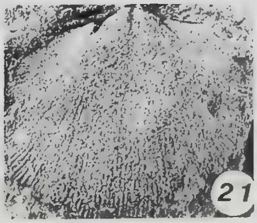
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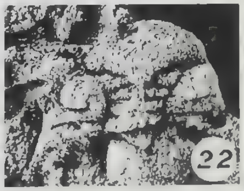
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To date, the High Mains Formation contains a fauna of thirteen brachiopod (Harper 1981) and five trilobite (Owen 1986) taxa. The brachiopods are characterized by a relative abundance of *Hirnantia sagittifera* (M'Coy), *Eostropheodonta* aff. *hirnantensis* (M'Coy) and *Hindella crassa* (J. de C. Sowerby) *incipiens* (Williams), important elements of the terminal Ordovician *Hirnantia* fauna, and less common *Glyptorthis*, *Plaesiomys*, *Platystrophia*, *Eochonetes*, *Eopholidostrophia*, *Fardenia*, *Rostricellula*, *Hypsiptycha* and *Eospirigerina* and an indeterminate enteletacean. With the exception of *Hypsiptycha*, all these forms have congeners in the underlying Drummuck Group. Moreover small individuals of *H. crassa incipiens* have been described previously from the Ladyburn Starfish Beds within the upper Rawtheyan South Threave Formation near the summit of the Drummuck Group (Reed 1917: 955; pl. 24, fig. 55) whilst Mitchell (1977: 54) has described and figured a species of *Hirnantia* from the Cautleyan Killey Bridge Formation, which is along strike in the Pomeroy inlier of the north of Ireland.

The Girvan fauna is quite distinct from other *Hirnantia* faunas (cf. Rong 1984*a*); whilst the fauna is dominated by key members of the *Hirnantia* fauna, it is of moderate diversity and supplemented by essentially relict North American forms. It is nevertheless different from other coeval assemblages, for example the *Holorhynchus* and Older Edgewood faunas (Rong 1984*b*: 117). Similarly, the trilobite fauna is dominated by North American relicts (Ingham *in* Harper 1981; Owen 1986).

The succeeding Mulloch Hill Conglomerate unconformably overlies the Drummuck Group. This formation is dominated by units of polymict, poorly sorted, of either clast- or matrix-supported conglomerate. The clasts range in diameter from a few centimeters up to 15 cm; a variety of lithologies is represented as is a range of shapes from near rounded to angular. The conglomerate units are separated by thinner beds of coarse impure quartz sandstone which are locally fossiliferous. Cocks & Toghill (1973) considered a shallow water environment of deposition likely for the unit whilst more recently Walton (1983: 133) indicated the sedimentology and fauna of the formation to be suggestive of shallow, shelf conditions. The available data however suggest an equally feasible alternative. The nature and thickness of the formation, in excess of 100 m, together with an ability to cut through some 350 m of strata over a distance of about five miles, suggest the Mulloch Hill Conglomerate was deposited in a channel across a gradient of depths. Clearly in the vicinity of Girvan the unconformity was not subaerial but rather resulted from downslope channelling during the earliest Silurian (see also Ingham 1978).

The fauna of the Mulloch Hill Conglomerate, although locally abundant within the sandstone units, is of low diversity. It is dominated by crinoid ossicles and the brachiopods *Cryptothyrella angustifrons* (Salter) and a species of *Rhynchotreta* (Cocks & Togill 1973). Both species have near identical relatives in the fauna of the upper Rawtheyan Ladyburn Starfish Beds (Harper 1979*a*). Such associations characterize shallow water environments created during the early Llandovery global transgression (Sheehan 1977).

Discussion

The faunal succession across the Ordovician–Silurian junctions indicates three phases of development: (a) above the Rawtheyan–Hirnantian transition a marked decrease in diversity concomitant with the development of a fauna comprising relict middle Ashgill elements of the North American province together with more abundant key taxa of the *Hirnantia* fauna, (b) during the early and middle Rhuddanian very low diversity faunas characteristic of the, then, recently colonized shallow water environments in the North American province, and (c) the arrival during the middle and late Rhuddanian of diverse, more typically Llandovery, shelly faunas. The former two events are accompanied by channel development during the regression whilst the latter is concomitant with net transgression. Similarly in the more complete and stable boundary section of the Oslo Basin relict Ordovician forms are not displaced by more typical Silurian elements until at least the middle Rhuddanian (Baarli & Harper 1986).

The mutual relationships of the basal Silurian facies and their southwestward overlap and overstep have been rationalized recently by Bluck (1983: fig. 6). Such features are considered to be the result of deposition on blocks of Ordovician strata separated by high-angle listric faults with approximately east to west trends. Evidence of fault-controlled sedimentation has been documented within the middle Ordovician succession of the Girvan district in the classic study by Williams (1962), more recently refined by Ince (1984). Whilst the disposition and relative movement of such blocks can at least partly explain lower Silurian facies patterns in the Girvan district, a mechanism is available also to provide substantial and continued local regressions during the late Ordovician and early Silurian. The relative downfaulting of sequential blocks to the south, during extensional phases, may have resulted in the rotation of each block about an axis parallel to the trend of the listric faults; consequently the leading apex of each block may have become emergent. The overall effect locally is one of regression and channel development across relatively steep slopes. Both faunal and facies development thus occurred in a tectonically active environment at Girvan, against a background of global regression and transgression during the late Ashgill and early Llandovery respectively.

Acknowledgements

I thank Dr D. M. Williams for advice regarding sedimentology and for his comments on the manuscript. Dr A. W. Owen kindly provided unpublished data on the High Mains trilobites and valuable discussion. Much of the fieldwork was carried out during tenure of a N.E.R.C. research studentship at Queen's University, Belfast.

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Base of the Silurian in the Lake District and Howgill Fells, Northern England

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Synopsis

The basal Silurian in the Lake District and Howgill Fells is divided into four slightly different depositional zones, only one of which shows a provable base to the *acuminatus* Zone, being underlain by a *persculptus* fauna and overlain by an *atavus* fauna. Other sections have 'Basal Beds' which certainly represent very condensed deposition of carbonates, perhaps involving non-sequences. The varied environments are interpreted as part of an offshore fault-scarp-cum-ridge-and-hollow system paralleling the Iapetus Suture and situated upon the southern (northward-dipping) plate.

There are essentially four rather different depositional environments at the Ordovician–Silurian boundary in the Lake District proper and in the Howgill Fells; and these are each different again from the facies and faunal development at Cross Fell, dealt with by Wright elsewhere in this volume. The four types are shown in Figs 1–4: although drawn diagrammatically it is important to realize that there are no exposure gaps in the region of the boundary, and that the sections in the Howgill Fells and western Lake District (Figs 1, 4) can be confirmed in many other nearby sections.

The *acuminatus* Zone fauna, the new basal Silurian zone, is well represented except in one small region only, namely the classic Skelgill section (Fig. 3), the type section of the Skelgill Beds black shale formation. On this section there is a thin, hard, partly calcareous and partly siliceous shelly mudstone (usually referred to in the literature as the Basal Beds). A similar bed occurs in the Howgill Fells, but the age on Skelgill could range from the *persculptus* Zone to the lower *atavus* Zone inclusive, for it is underlain by Ashgill Shales (Hirnantian; and probably of *anceps* Zone age) and overlain by upper *atavus* Zone black shales. The Basal Beds certainly represent a period of condensed deposition and possibly of non-sequence. There is no direct evidence of hardground criteria. The shelly fossils include *Atrypa flexuosa* and may represent relatively deep water community life with low diversity.

In the Howgill Fells and the eastern Lake District (Figs 1, 2) the *acuminatus* Zone is well established but its base, and hence the base of the Silurian, cannot be proved: the Basal Beds in the Howgill Fells might be of *persculptus* Zone age, but a possible bentonite separates those beds from the thin *acuminatus* Zone black shale; and at Browgill a 0·08 m rottenstone, possibly the lithological and stratigraphical equivalent of the Basal Beds, separates Hirnantian Ashgill shales from black, *acuminatus* Zone shales.

Only in the western Lake District (Fig. 4) can the base of the Silurian be unequivocally placed, albeit on numerous sections in the region. The Yewdale Beck section is well and continuously exposed, and above 0·3 m of beds with a good *persculptus* Zone fauna are 11 m of black shales with a very rich assemblage of *acuminatus* Zone graptolites (Hutt 1974). The *persculptus* Zone also contains numerous shelly fossils of most groups, but they have not been extensively studied. The Ashgill Shales below them yield numerous brachiopods and rarer trilobites giving a Hirnantian age to the Ashgill Shales, but graptolites in these beds are rare. The *acuminatus* Zone black shales yield shelly fossils only very infrequently and none to date have proved to be of diagnostic value. In every other respect, however, the Yewdale Beck section provides a good confirmatory section for the base of the Silurian, especially as an almost infinite number of both natural and artificial sections are available in the general region of Coniston and on the fells and streams to the southwest of that town. Graptolites from these sections can be collected by the hundred and, as with all other *acuminatus* Zone faunas mentioned above, almost all the typical species of the zonal assemblage occur.

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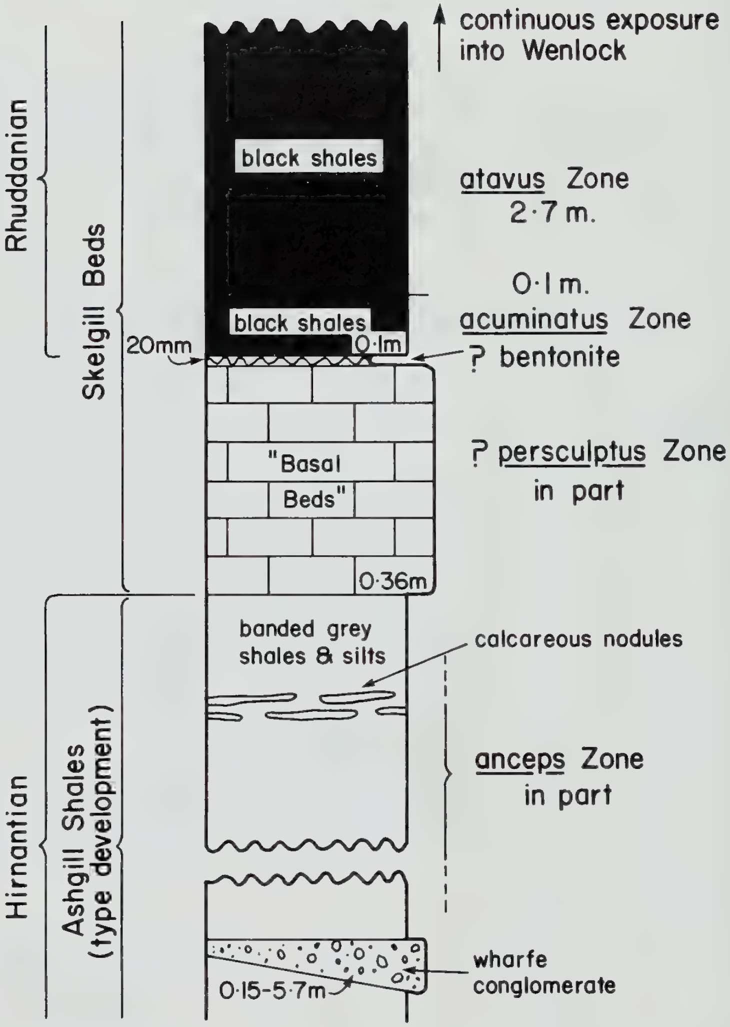
continuous exposure into Wenlock
black shales
atavus Zone 2·7 m.
0·1 m.
acuminatus Zone
black shales
0·1m
20mm
? bentonite
"Basal Beds"
? persculptus Zone in part
0·36m
banded grey shales & silts
calcareous nodules
anceps Zone in part
wharfe conglomerate
0·15-5·7m
Rhuddanian
Skelgill Beds
Hirnantian
Ashgill Shales (type development)

Fig. 1 Howgill Fells: beds about the Ordovician–Silurian boundary on Spengill, Grid Reference SD 698998.

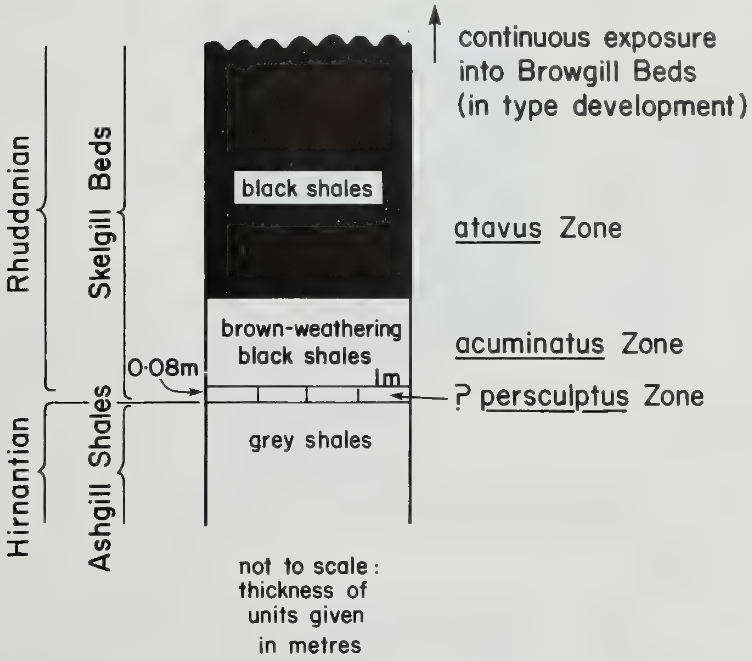
Rhuddanian
Skelgill Beds
Hirnantian
Ashgill Shales
continuous exposure into Browgill Beds (in type development)
black shales
atavus Zone
brown-weathering black shales
1m
acuminatus Zone
0·08m
? persculptus Zone
grey shales
not to scale: thickness of units given in metres

Fig. 2 Eastern Lake District: beds about the Ordovician–Silurian boundary on Browgill, NY 4974 0587.

Rickards (1978) attempted a general interpretation of the environment of deposition of the basal Silurian strata, envisaging a west- or northwest-facing fault scarp, according to Hutt (1974) active during deposition of the early Llandovery, against which were deposited deeper offshore, black shales and upon which and behind which were deposited the Basal Beds and their equivalents. By upper *atavus* Zone times the scarp feature was further submerged and covered in black shale deposition. Associated with these features were a series of ridges and hollows striking ENE/WSW, that is roughly the same as the fault scarp strike. The hollows received a greater thickness of black shale in a more highly anaerobic environment (Rickards 1964). The ridge and hollow system persisted in the Howgill Fells region, and possibly in the main Lake District outcrop, until late in the Llandovery.

Thus the onset of the Silurian in the Lake District is marked by condensed deposition of shelly limestone, and possible non-sequences, in the eastern, presumed shoreward or shallower region; and by relatively thick, black shale deposition in the western Lake District. The post-glacial marine transgression is recorded in the gradual spread of black shale deposition over the whole region, the last area to succumb being the eastern Lake District area of Skelgill which is interpreted as being on the crest of an old scarp structure, itself certainly operative as far back as the Caradoc. It seems likely that the region was situated atop the northward-dipping plate, south of the Iapetus Suture. The scarp and ridge/hollow systems may be a result of the northwards subduction process, to which they are parallel, and which resulted in a combination of compressional and extensional features.

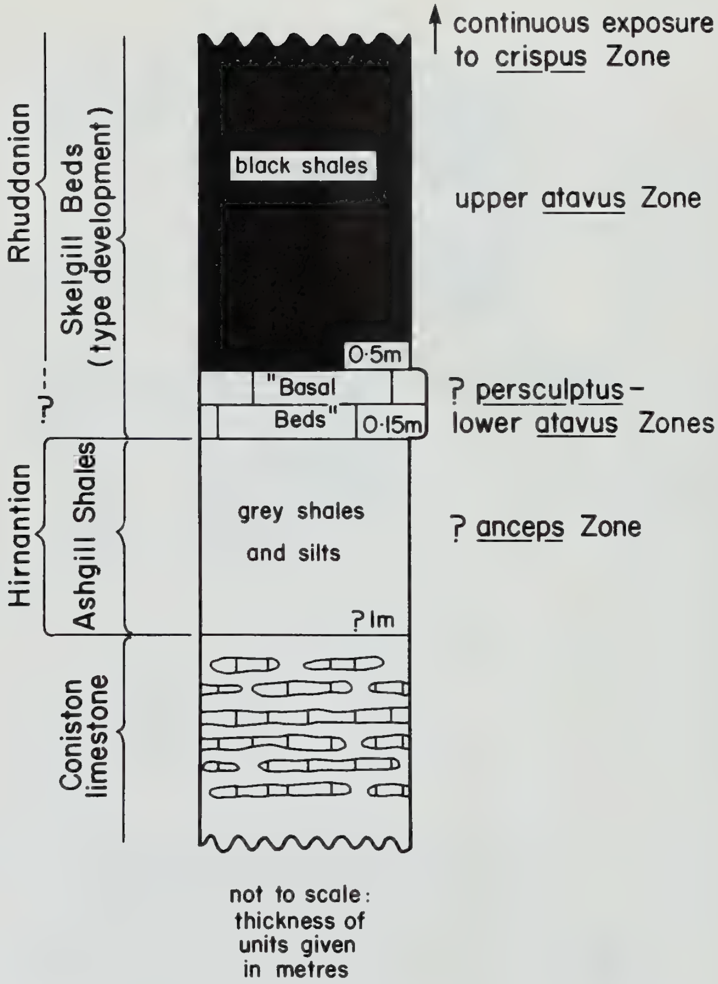
continuous exposure to crispus Zone
upper atavus Zone
? persculptus – lower atavus Zones
? anceps Zone
black shales
0·5m
"Basal Beds"
0·15m
grey shales and silts
? 1m
Rhuddanian
?
Hirnantian
Skelgill Beds (type development)
Ashgill Shales
Coniston limestone
not to scale: thickness of units given in metres

Fig. 3 Eastern Lake District: beds about the Ordovician–Silurian boundary on Skelgill, NY 3964 0320.

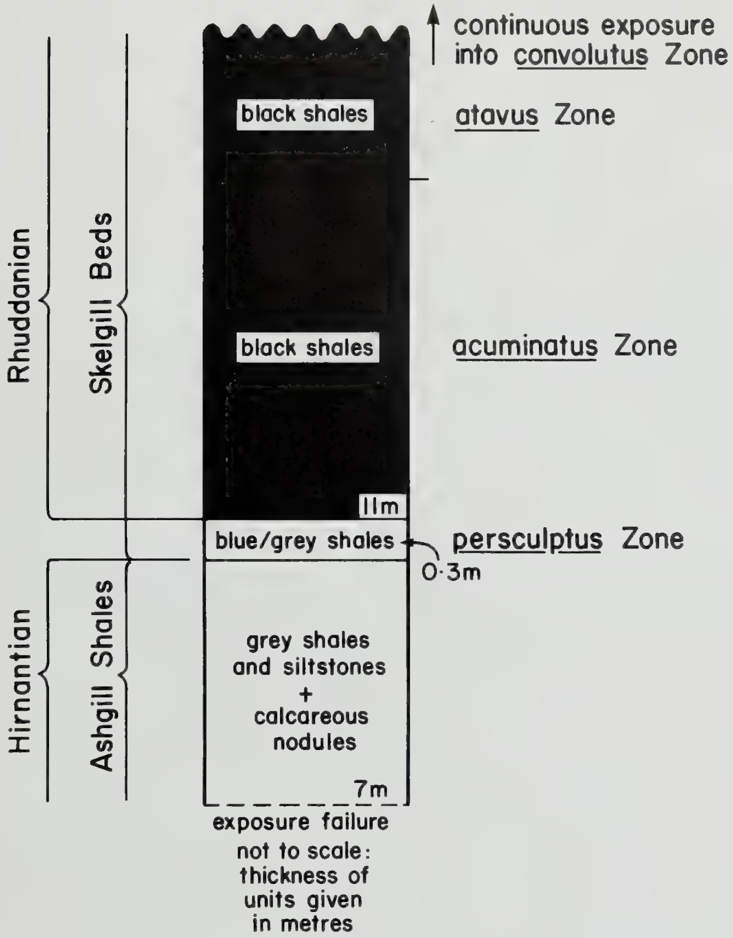
continuous exposure into convolutus Zone
black shales
atavus Zone
black shales
acuminatus Zone
11m
blue/grey shales
persculptus Zone
0·3m
grey shales and siltstones + calcareous nodules
7m
exposure failure
not to scale: thickness of units given in metres
Rhuddanian
Skelgill Beds
Hirnantian
Ashgill Shales

Fig. 4 Western Lake District: beds about the Ordovician–Silurian boundary at Yewdale Beck, SD 3073 9858.

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The Ordovician–Silurian boundary at Keisley, Cumbria

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Synopsis

At Keisley, in the Cross Fell Inlier of Cumbria, the lowest Silurian graptolite biozone recorded until recently was that of *A. atavus*, with the topmost of the underlying carbonates regarded as being of either Lower Llandovery or Hirnantian age. A temporary excavation has confirmed the Hirnantian age of the latter, and with the discovery in the overlying clastic sediments of the biozones of both *G. persculptus* and *P. acuminatus*, the Ordovician–Silurian boundary is now accurately located.

Although Upper Ordovician and Lower Silurian rocks crop out in the Cross Fell Inlier of northern England, the area is much faulted (Shotton 1935). Moreover, where reasonably continuous graptolite sequences of the Lower Silurian are exposed in Swindale Beck (Knock) and in Great Rundale Beck (Marr & Nicholson 1888: 699; Burgess & Rushton 1979: 23), the lowest biozones (below *Coronograptus cyphus*) are missing. Until recently, the earliest Silurian graptolite biozone was that recorded from the road cutting to Keisley Quarry by Marr (1906: 485) and reported by him as indicating the *Dimorphograptus confertus* Zone of Marr & Nicholson (1888). The lowest part of that zone has been shown by Rickards (1970) to equate with the *Atavograptus atavus* Biozone, and the presence of beds of this age was confirmed by Rickards from graptolite material excavated in 1965 from this locality by Temple (1968: 2).

On the Upper Ordovician side of the boundary the stratigraphical relationships and precise age of the main unit, the Keisley Limestone, have been debated for many years. The limestone has been a source of geological interest since the last century as it contains a prolific shelly fauna, is of distinctive lithology, and has a peculiar morphological form referred to as a 'knoll' by Marr (1906: 485). The views on various aspects of this mudmound have been discussed by Wright (1985); only the relationships of the carbonate mudmound to the *atavus* Biozone graptolite shales are relevant in the present context.

Marr (1906: 485) noted that the Ashgill Shales, which do occur in Swindale Beck, were not present at Keisley; and as there was insufficient room for these beds between the Silurian graptolite shales and the nearest outcrops of Keisley Limestone, he interpreted the junction as a faulted one. Burgess (1968: 343) noted that along the track leading to the quarry, the massive limestone was succeeded by calcareous mudstones and limestone nodules which were in turn overlain by the graptolite shales 'in apparently conformable sequence', and the presence of this apparently unfaulted and conformable relationship was subsequently reiterated by Burgess *et al.* (1970: 170), despite the discontinuous nature of the outcrops. An extensive brachiopod and trilobite fauna was collected by Temple (1968, 1969) from weathered limestone bands associated with unfossiliferous shales at the bend in the quarry track; this outcrop was separated by a few metres from those of both the underlying massive limestone and the overlying *atavus* Biozone shales, and the extensive fauna interpreted by Temple as being of Lower Llandovery age, a view supported by Burgess & Rushton (1979: 23) but not by Ingham & Wright (1972: 47), who regarded it as being of Hirnantian age.

The difficulty with the Keisley locality is that the beds immediately below the established *atavus* Biozone graptolite shales are concealed beneath the trackway to the quarry. To overcome this a temporary trench was dug with the aid of a mechanical digger and the complete sequence exposed (Wright 1985). Fig. 1 shows the position of the trench across the trackway and Fig. 2 the lithological log obtained.

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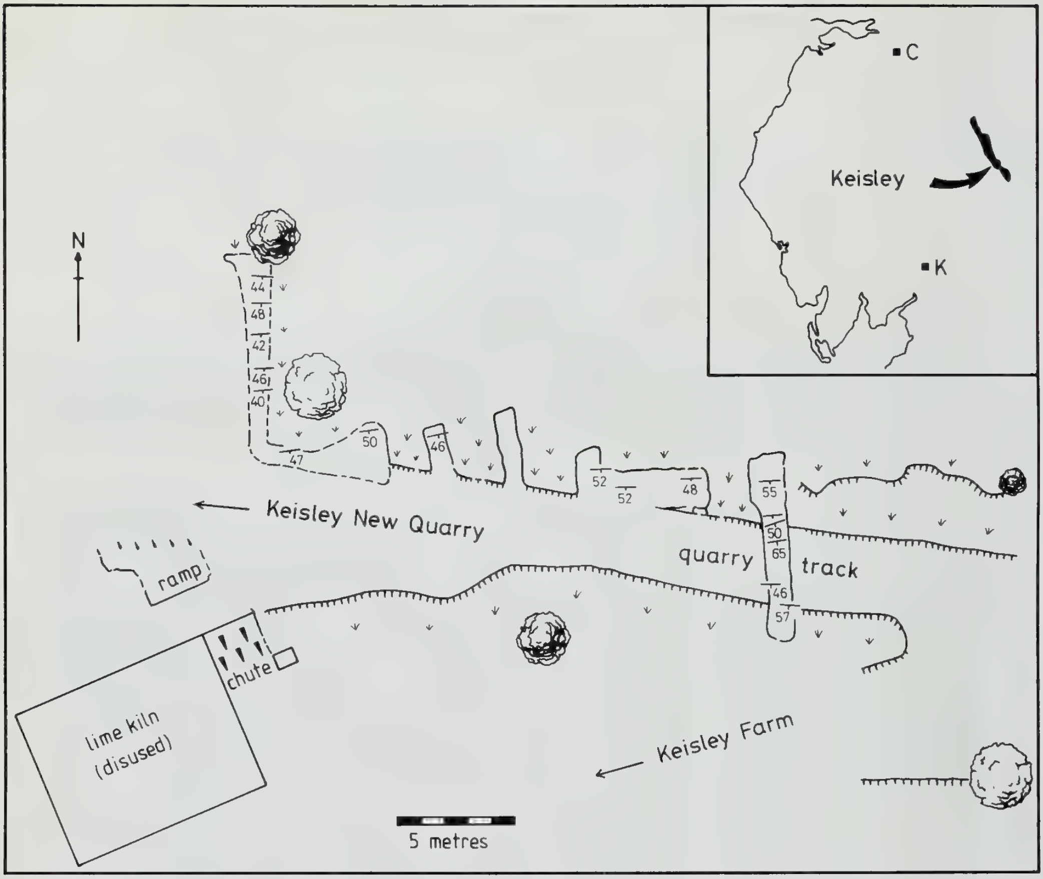

Fig. 1 Plan showing the position of the temporary trench excavated across the trackway at the eastern end of Keisley New Quarry to reveal the Ordovician–Silurian boundary (National Grid Ref. NY 7137 2379). The strikes in the trench were taken on the trench floor except for the two strikes at the southern end which are in the trench walls and thus well up in the *atavus* Biozone. Stippling in the bank to the east of the trench indicates outcrops of fossiliferous weathered limestones, the fauna of which was described by Temple (1968, 1969). Stylized trees (not to scale) represent two ash (light outlines), two sycamores (dark outlines) and a hawthorn (small figure). The inset figure shows the position of Keisley in the Cross Fell Inlier (shaded) in relation to north-west England (C—Carlisle; K—Kendal).

The lower part of the sequence up to and including unit 8 (numbering as in Wright 1985) consists of alternations of bedded limestones or calcareous nodules with calcareous siltstones. The bedded bioclastic limestones are fresh and although pelmatozoan debris, bryozoan fragments and the occasional brachiopod (including *Hirnantia sagittifera*) are to be seen on the bed surfaces, faunal lists are scant compared with those of Temple (1968) obtained from the well weathered material above the trackway. Gastropods, ostracodes and a few trilobites have been observed in thin sections of the trench limestones in which abundant *Girvanella* is probably the most revealing element palaeoenvironmentally.

The unit 7 siltstone, while by no means abundantly fossiliferous, does have a shelly fauna in the form of moulds, albeit in a broken and fragmented state. The diverse fauna includes the brachiopods *Dolerorthis praeclara*, *Hindella* sp., *Hirnantia sagittifera*, ? *Oxoplecia*, *Paracraniops* sp., *Reuschella inexpectata*, *Skenidioides scoliodus*, *Sphenotreta* sp. and *Toxorthis proteus*

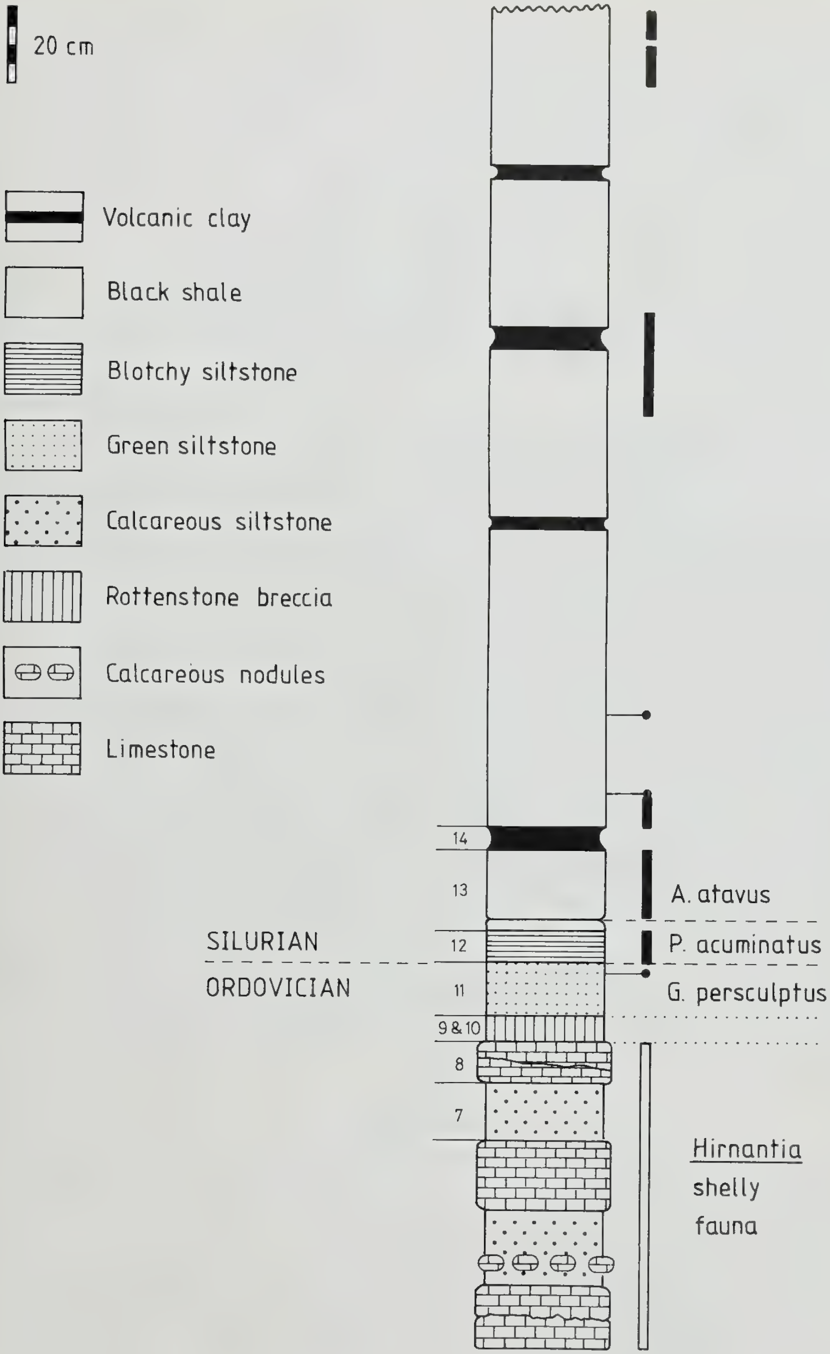
20 cm
Volcanic clay
Black shale
Blotchy siltstone
Green siltstone
Calcareous siltstone
Rottenstone breccia
Calcareous nodules
Limestone
14
13
12
11
9 & 10
8
7
SILURIAN
ORDOVICIAN
A. atavus
P. acuminatus
G. persculptus
Hirnantia
shelly
fauna

Fig. 2 The lithological log obtained for the trench cutting across the quarry track of Fig. 1, showing the position of the Ordovician–Silurian boundary. Numbers of lithological units discussed in the text are as in Wright (1985). The black spots and bars to the right of the log respectively indicate specific horizons or bulk samples yielding graptolite assemblages.

together with dalmanellid, lingulide, orthid, sowerbyellid, strophomenide and triplesiid fragments. In addition to pelmatozoan and bryozoan debris, trilobite, bivalve and hyolith fragments are also recorded. This is a *Hirnantia* shelly fauna, and differs principally from Temple's fauna in the apparent complete absence of craniids which accounted for more than two-fifths of the entire brachiopod assemblages from the weathered limestones (Temple 1968: 9).

Overlying these beds is a thin (7 cm) rottenstone breccia (units 9 and 10). This is the only indication of a break in the sequence and is interpreted as the result of minor tectonic movement along the surface of lithological change from the underlying carbonate dominated sequence to the overlying fine-grained and non-carbonate clastics. Angular clasts of both fossiliferous shelly Hirnantian and unfossiliferous greenish siltstone (matching the unit 11 sediment) occur in the breccia. No diagnostic shelly fossils have been located in the sequence above unit 10. The first graptolites recovered by Rickards are from a horizon 2 cm below the top of unit 11 and indicate the *Glyptograptus persculptus* Biozone. This fauna comprises *Climacograptus* cf. *miserabilis*, *Climacograptus* ? *medius*, *Glyptograptus* sp. and *Glyptograptus* ex gr. *persculptus*.

Unit 12 is an 8 cm unit of silt with a blotchy and mottled appearance produced by an increase in the proportion of dark muddy silt that first appears in the greenish siltstones of unit 11 (Wright 1985: 269). Despite clear evidence of bioturbation, a small graptolite fauna from a bulk sample of the unit contained specimens of *Climacograptus normalis* and cf. *Parakidograptus acuminatus*, and indicates the presence of the *Parakidograptus acuminatus* Biozone. The Ordovician–Silurian boundary at Keisley is accordingly placed at the base of lithological unit 12. This seems to be the most logical horizon although, as noted previously (Wright 1985), there is clearly a little uncertainty regarding the precise appearance of the *acuminatus* fauna within a bulk sample taken from the 8 cm unit.

Unit 13 lithologically shows a further stage in the transition from the greenish siltstones at the base of unit 11 towards the micaceous black silty shales of the overlying sequence. In this unit the dark material is dominant, although some horizons and patches of the greenish-grey siltstones still occur; concomitantly with the overall colour change, bioturbation disappears. At 2·5 cm above the base of this unit, the first of a series of bentonite clays occurs. A fauna collected from a bulk sample above this clay (Fig. 2) yielded *Climacograptus medius*, *Climacograptus* cf. *normalis* and *Dimorphograptus* sp. This assemblage is identified by Rickards as a post-*acuminatus* one, i.e. from the base of the *atavus* Zone. Accordingly the *acuminatus–atavus* boundary is placed at the thin bentonite band, which is a useful marker that may assist with correlation elsewhere, although the appearance of *atavus* Biozone bentonites in the Keisley trench is a major surprise in the northern England context (Wright 1985). The increasingly rich graptolite faunas from the overlying sequence in the trench all belong to the *atavus* Biozone.

Thus although the *persculptus* and *acuminatus* Biozones occur in thin lithological units at Keisley, both do occur and accordingly enable the Ordovician–Silurian boundary to be precisely located.

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Ordovician–Silurian boundary strata in Wales

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Synopsis

Ordovician–Silurian boundary strata in Wales belong to the shelly facies in the south and east, and to the graptolitic facies in the north and west. In the graptolitic facies the zones of *Dicellograptus anceps*, *Glyptograptus persculptus* and *Parakidograptus acuminatus* occur, but the *Climacograptus*? *extraordinarius* Zone is not known. The *anceps* Zone is restricted to central and west Wales; the *persculptus* Zone is widespread and is preceded by a sudden lithological change; the *acuminatus* Zone is preceded by a more gradual lithological change. Graptolites occur sporadically in boundary strata of the shelly facies but are not abundant enough for the base of the *acuminatus* Zone to be recognized in this facies. Records of the *Hirnantia* fauna in Wales are summarized.

Introduction

As a result of Caledonian and Hercynian folding the Ordovician–Silurian boundary strata in Wales form a complex arcuate pattern striking approximately NE–SW through much of central Wales but becoming east–west in south-west Wales and SE–NW in north-east Wales. The length of outcrop is approximately 750 km. The outcrop is shown in Fig. 1, together with index numbers by which individual areas and the references relating to them are cited in the text.

In places on the outward (S, SE or E) side of the Caledonian fold belt in Wales, as in the adjoining parts of England, the local base of the Silurian is formed by late Llandovery (post-*convolutus* or post-*sedgwickii*) or Wenlock strata transgressive onto pre-Ashgill strata. This relation is found in the southernmost outcrop (but not in the main northern outcrop) at Haverfordwest (1a), near Llandeilo (2), from north of Llandovery (4) to Garth (5a, b), near Builth Wells (6), east of Abbey-Cwmhir (7), and east of Welshpool (25, 26). Flanking this marginal area of late Llandovery transgression there is an unconformity of lesser magnitude between the early Llandovery and the Ashgill (and Caradoc) near Welshpool (27) and Llansantffraid ym Mechain (31), and although the gap continues to diminish northwards and westwards it is recorded as still present in the Meifod and Vyrnwy areas (28, 29). Elsewhere in Wales the early Llandovery is believed to follow the topmost Ordovician with no sedimentary gap.

Boundary strata

Ordovician–Silurian boundary strata in Wales show two facies, shelly and graptolitic. The shelly facies consists of detrital sediments, mainly of the silt and sand grades, with a fauna predominantly of brachiopods. The graptolitic facies consists of fine detrital sediments (mudstones and shales) with some coarser horizons interpreted as turbidites, and with a fauna almost exclusively of graptolites.

In pre-*persculptus* Zone strata the dichotomy into shelly and graptolitic facies is not as clearly defined as later. The strictly graptolitic facies, as defined by the recorded presence of the *Dicellograptus anceps* Zone, is much more restricted in occurrence (to central and west Wales—16, 18, 19, 20) than the *persculptus* Zone, and even where both zones occur in the same area the intervening strata are either unfossiliferous (16, 18, 19) or include shelly fossils (20). Along the outcrop north-west of the Towy anticline (8–14), for instance, where the *persculptus* Zone is graptolitic, the very thick underlying strata yield only sporadic graptolites (not diagnostic of the *anceps* Zone), being otherwise unfossiliferous or with a few shelly fossils. The restriction of the demonstrable *anceps* Zone to central and west Wales and the wider extent eastwards of the *persculptus* and *acuminatus* Zones are consistent with regression during *anceps* Zone time followed by transgression during the *persculptus* Zone. The *extraordinarius* Zone has not been

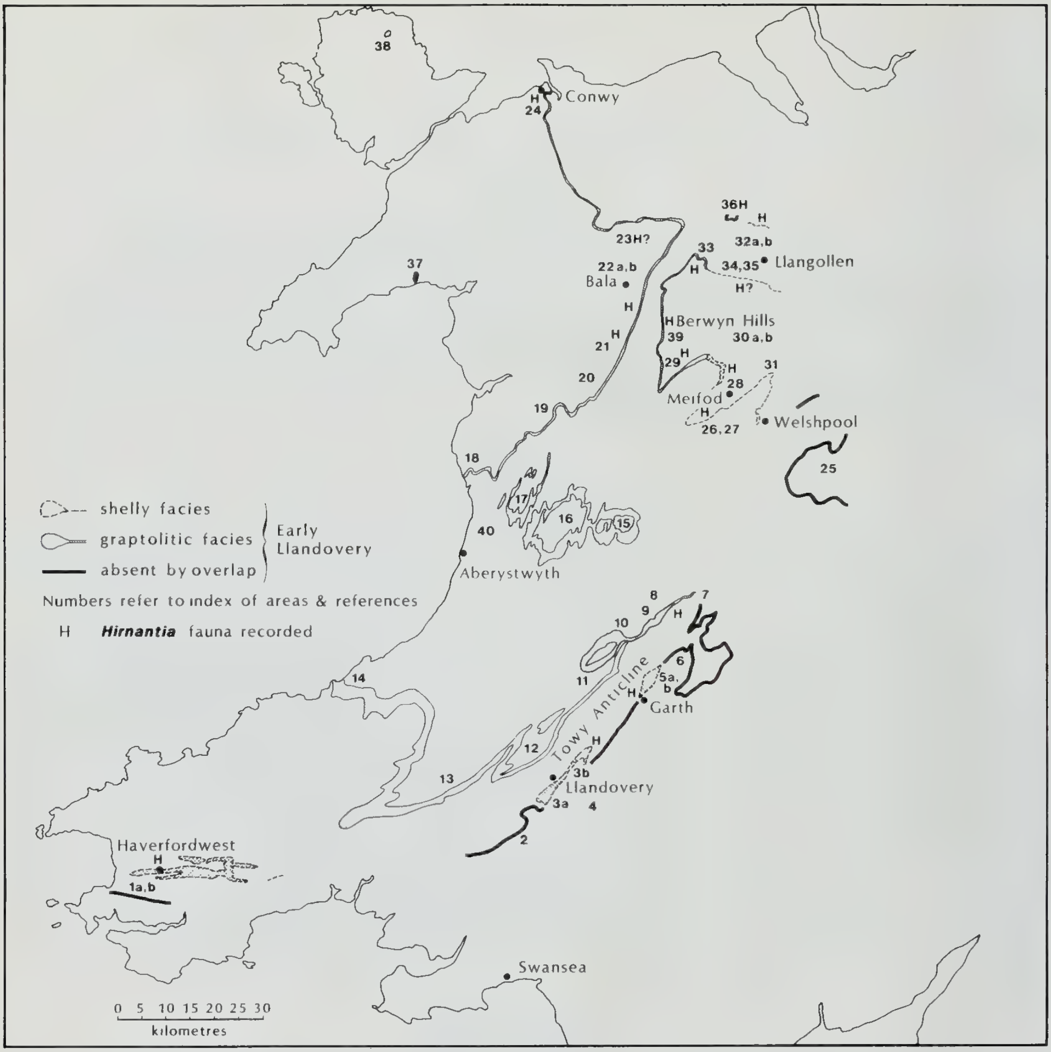
38
Conwy
H
24
36H
H
23H?
33
32a,b
22a,b
Bala
H
34,35
Llangollen
H?
37
H
21
H
Berwyn Hills
39
30a,b
H
29
H
31
28
20
Meifod
H
19
26,27
Welshpool
18
25
17
16
15
40
Aberystwyth
shelly facies
graptolitic facies
absent by overlap
Early Llandovery
Numbers refer to index of areas & references
H *Hirnantia* fauna recorded
8
7
9
H
10
6
11
5a,b
H
Garth
Towy Anticline
H
12
13
3b
Llandovery
3a
4
14
2
Haverfordwest
H
1a,b
Swansea
0 5 10 15 20 25 30
kilometres

Fig. 1 Ordovician–Silurian boundary outcrop areas in Wales and the Welsh Borderland. 1, Haverfordwest: 1a, Strahan *et al.* 1914; 1b, Cocks & Price 1975. 2, Llandeilo, Williams 1953. 3, 4, Llandovery: 3a, Jones 1925; 3b, Jones 1949; 4, Cocks *et al.* 1984. 5, Garth: 5a, Andrew 1925; 5b, Williams & Wright 1981. 6, Builth Wells, Jones 1947. 7, Abbey-Cwmhir, Roberts 1929. 8, 9, Rhayader: 8, Lapworth 1900; 9, Kelling & Woollands 1969. 10, Rhayader to Abergwesyn, Davies 1928. 11, Abergwesyn to Drygarn, Davies 1926. 12, Pumpsaint, Davies 1933. 13, Llansawel, Drew & Slater 1910. 14, Llangranog, Hendricks 1926. 15, Llanidloes, Jones 1945. 16, Plynlimon, Jones 1909. 17, Machynlleth, Jones & Pugh 1916. 18, Towyn and Abergynolwyn, Jehu 1926. 19, Corris, Pugh 1923. 20, Dinas Mawddwy, Pugh 1928. 21, Llanuwchllyn-Llanymawddwy, Pugh 1929. 22, Bala: 22a, Elles 1922; 22b, Bassett *et al.* 1966. 23, Cerrigydrudion, Marr 1880. 24, Conwy, Elles 1909. 25, Shelve area, Whittard 1932. 26, 27, Welshpool: 26, Wade 1911; 27, Cave 1965. 28, Meifod, King 1928. 29, Lake Vyrnwy, King 1923. 30, Berwyns: 30a, Wedd *et al.* 1929; 30b, Brenchley & Cullen 1984. 31, Llansantffraid ym Mechain, Whittington 1938. 32, Llangollen: 32a, Groom & Lake 1908; 32b, Hiller 1980. 33, Corwen, Lake & Groom 1893. 34, Llangollen, Wills & Smith 1922. 35, Llangollen, Wedd *et al.* 1927. 36, Mynydd Cricor, Smith 1935. 37, Criccieth, Roberts 1967. 38, Anglesey, Greenly 1919. 39, W. Berwyn, A. W. A. Rushton & J. T. Temple (unpublished). 40, Aberystwyth and Machynlleth, Cave & Hains 1986.

recognized in Wales, but there is ample room for it: the barren strata between the *anceps* and *persculptus* Zones in areas 16, 18, 19, 20 are respectively 730 m, 1000 m, 690 m, and 180 m thick.

In the *persculptus* Zone and the succeeding early Llandovery the dichotomy into shelly and graptolitic facies is well shown. The shelly facies forms a narrow belt running through Haverfordwest (1a, b), the Llandovery (3a, b, 4) and Garth (5a, b) areas (which form north-westward salients from the adjacent line of outcrop along which the strata are missing), and the eastern end of the Berwyn dome (27–32, 34–36). The transition from shelly to graptolitic facies of the *persculptus* Zone and early Llandovery takes place in south and central Wales across the Towy anticline (between for instance Llandovery [3a, b, 4] and Pumpsaint [12]), and in north-east Wales probably north-westwards across the Berwyn dome. The *persculptus* and *acuminatus* Zones are widespread, having been recorded from north-west of the Towy anticline (10–12) as well as through most of central and west Wales (14–21, 40). *G. persculptus* occurs on the western outcrop around the Berwyn Hills at Nant Pant-y-llidiart, north of Lake Vyrnwy (39), and there is an informal record of the species at Bwlch yr Hwch, 5 km SE of Bala (Jones *in* Pugh, 1929: 274–5). The *persculptus* Zone (but not the *acuminatus* Zone) has also been recorded from the north end of the Towy anticline (7, 9), and *G.* cf. *persculptus* occurs at Garth (5a). The early Llandovery graptolite succession between Bala (22a, b) and Conwy (24) is still in need of reinvestigation. In the two small isolated outliers near Criccieth (37) and in Anglesey (38) the early Llandovery is in graptolite facies, but in both cases the relationship to the Ordovician is obscure and neither the *persculptus* nor the *acuminatus* Zones are recorded.

A sudden and striking lithological change heralds the incoming of *persculptus* Zone graptolites in west and central Wales (14–20, 40): the underlying strata are very thick, usually unfossiliferous, often unbedded, well cleaved or doubly cleaved, and with many 'grit' bands; the *persculptus* Zone strata (the 'Mottled Beds') are mudstones 5–25 m thick, well-bedded, often with mottled pale bands (interpreted as bioturbated—Cave & Hains 1986) and with a thin band crowded with the zone fossil about 1 m above the base. The suddenness of the lithological change preceding the appearance of *G. persculptus* in this part of Wales betokens some physical change in the conditions of deposition, and this evidence also is consistent with a *persculptus* transgression following regression. A similar lithological contrast at this horizon is also found north of the Towy anticline (9–11), although not strongly marked in the south of the outcrop (12).

There is also a lithological change below the *acuminatus* Zone in west and central Wales (15–20, 40), but it is more gradual than that below the *persculptus* Zone, the hard resistant mottled mudstones of the latter zone being gradually replaced by rusty red- and yellow-weathering mudstones without bioturbation (40). A similar change occurs at this horizon north of the Towy anticline (10–12). In both areas the change probably precedes the end of the *persculptus* Zone (40, 12).

Hirnantia fauna in Wales

Around the Berwyn dome and near Llangollen there are developed 'grits' which have been taken as either topmost Ordovician (35) or basal Silurian (28, 29): Craig-wen Sandstone (28), *Meristina crassa* Sandstone (29), Allt-gôch Grit (30), Corwen Grit (33), Glyn Grit (32), Plâs uchaf Grit (35). These grits have been interpreted as channel-fill deposits formed during the Hirnantian regression (30b). 'Grits', possibly of the same age as those around the Berwyns, also occur in the north and east of the Bala area (Calettwr Quartzite—22b) and along the little-known outcrops north of Bala, i.e. at Cerrigydrudion (23) and Conwy (Conwy Castle Grit—24). South of the Berwyns there are 'grit' bands near Abbey-Cwmhir (7) which are mapped as topmost Ordovician but whose relationship to the *persculptus* Zone strata occurring about 3 km to the west needs reinvestigation.

Many of the 'grits' in these different areas include elements of the *Hirnantia* fauna (Fig. 1), for which Brenchley & Cullen (1984: 122) give faunal lists at various Welsh localities. To these authors' list for 'Meifod' (i.e. Craig-wen quarry, near Meifod) may be added the record of the tretaspid indet. discovered on the Silurian Subcommission excursion in 1979, although the presence of pebbles of underlying strata in the Craig-wen Sandstone suggests the possibility of this being a derived fossil. The *Hirnantia* fauna also occurs in Afon Tanat on the western outcrop of the Berwyn Hills (39). The *Hirnantia* fauna at its type area south of Bala (22a) was considered by Pugh (1929: 273) to be pre-*persculptus* in age although no single section (except Jones' record at Bwlch yr Hwch—see above—which awaits confirmation) shows the one fauna succeeding the other. Further southwestwards along the outcrop (beyond 20) in west and west-central Wales the *Hirnantia* fauna dies out while the *persculptus* Zone fauna becomes more clearly developed. South of the Towy anticline the *Hirnantia* fauna has been recorded from Garth (5a, b) apparently in association with *G.* cf. *persculptus* (Williams & Wright 1981: 38), and from Haverfordwest (1b) in the St Martin's Cemetery Beds (Cocks & Price 1975: 710) whose relations to the *persculptus* Zone are unknown. The *Hirnantia* fauna has also recently been found in the Llandovery area (4) where it is considered (Cocks *et al.* 1984: 144) to underlie strata probably representing the *persculptus* Zone.

At Conwy (24) the *Hirnantia* fauna is underlain, as in the English Lake District, by strata containing abundant *Dalmanitina* [*Mucronaspis* auctt.], and this relationship is found also at Bala (22a) and in the Llanuwchllyn-Llanymawddwy area to the south (21). The trilobite persists southwestwards along the outcrop, as the facies change and the rocks thicken, even further (20, ?19) than the *Hirnantia* fauna. On the other hand the *Hirnantia* fauna around the Berwyns (28–30), at Abbey-Cwmhir (7) and at Garth (5) is not accompanied or preceded by *Dalmanitina* (except for a possible record in area 28—King 1928: 687), and although the absence of the latter trilobite in the Berwyns may be due to a stratigraphical gap below the *Hirnantia* 'grits', there is no evidence for such a gap at Abbey-Cwmhir or Garth, nor indeed at Llandovery where *Dalmanitina* is also absent. At Haverfordwest (1b) *Dalmanitina* occurs as part of an unusually rich *Hirnantia* fauna but is not found in underlying strata.

Descriptions of sections

Boundary strata of four areas merit description: Plynlimon-Machynlleth (16, 17, 40), where the sequence is graptolitic throughout and where the *persculptus* and *acuminatus* Zones are well developed; Llandovery (3a, b, 4) where the base of the Llandovery was originally defined; Haverfordwest (1a, b) and Garth (5a, b), in both of which there are apparently continuous successions in strata of predominantly shelly facies.

Plynlimon–Machynlleth (16, 17, 40). The succession in this area, which is wholly in the graptolitic facies, has recently been described in detail (Cave & Hains 1986). The best sections of the Mottled Mudstone Member are at the Cardiganshire Slate Quarry (National Grid ref. SN 6991 9595) and in a stream near Eisteddfa-Gurig (SN 7951 8409), but the faunal transition between the *persculptus* and *acuminatus* Zones has not been investigated in detail.

70–145 m	Cwmere Formation		Strata above Mottled Mudstone Member. Dark grey rusty-weathering mudstones: in middle, sandstones and siltstones near top of *acuminatus* Zone.
		5–25 m	Mottled Mudstone Member. Banded mudstone with pale bioturbated layers and phosphatic concretions (both disappearing in topmost 3 m). The lowest beds are unfossiliferous but about 1 m above the base is a thin layer (15–30 cm) with abundant *G. persculptus*. Pyritized *G. persculptus* also occur above this layer.
	Bryn-glâs Formation		Massive mudstone with splintery, phacoidal cleavage.

Garth (5b). The following section is obtained by mapping in strata of predominantly shelly facies near Garth, 32 km NE of Llandovery, Powys (Williams & Wright 1981).

250 m +	Sandstones & mudstones		Rhuddanian shelly fossils
77 m +	Garth Bank Formation		
11–51 m	Cwm Clŷd Formation		*Eostropheodonta hirnantensis*
0–30 m	Wenallt Formation	Speckly Sandstone Member	*Hirnantia* fauna. (Andrew [5a] records *G.* cf. *persculptus* and *Mesograptus* cf. *modestus parvulus* probably from this Member)
0–20 m		Ooid Member	*Hirnantia* fauna
0–65 m		Siltstones	*Brongniartella* cf. *robusta* (high Rawtheyan)

Llandovery (4). The following section (transect i_2 of Cocks *et al.* 1984) is exposed almost continuously along a forestry road in the north Llandovery area (base of section at Grid ref. SN 8467 3962). The *Hirnantia* fauna, however, is extrapolated from 1·3 km further west.

120 m	Bronydd Formation	Rhuddanian shelly fossils and graptolites suggesting *atavus* and *acinaces* Zones. Near base *Climacograptus normalis*
70 m	Scrâch Formation	(*Hirnantia* fauna in west)
—	Tridwr Formation	Rawtheyan shelly fossils and 'uppermost Ordovician' graptolites

Haverfordwest (1b). The following section (Cocks & Price 1975) is obtained by mapping in strata of predominantly shelly facies at Haverfordwest, Dyfed, but there are continuous exposures in road and railway sections upwards from about the middle of the Haverford Mudstone Formation (base of road section at Grid ref. SM 9573 1547).

85 m	Gasworks Sandstone Formation	Graptolites indicating *acinaces* or *atavus* Zones at top. Rhuddanian shelly fossils throughout
370 m	Haverford Mudstone Formation	Rhuddanian shelly fossils. *Climacograptus* cf. *normalis* near middle. ?*C. normalis* at 9 m above base
65 m	Portfield Formation	*Hirnantia* fauna at top, including Diplograptid undescr. sp.
—	Slade & Redhill Mudstone Formation	Rawtheyan shelly fossils

Conclusions

On the assumption (cf. Temple 1978) that graptolite zones are definable and recognizable entities, then because of the wide extent of the *persculptus* and *acuminatus* faunas in central Wales, the Ordovician–Silurian boundary defined beneath the *acuminatus* Zone is in principle widely applicable in Wales. It is not however directly applicable in the marginal belt characterized by boundary strata of shelly facies. Even in the recently reinvestigated Llandovery area (4), where there is an intermingling of shelly fossils and graptolites, the *persculptus* and *acuminatus* Zones are not firmly enough identified for the boundary to be recognized accurately.

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La Limite Ordovicien–Silurien en France

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Synopsis

The Ordovician and Silurian systems are well represented in France, but the boundary between them remains imprecise because there is generally a gap in the lower part of the Llandovery and/or the upper part of the Ordovician. The actual documentation for the Armorican Massif and the south-west of France is briefly revised.

Les systèmes ordovicien et silurien sont largement représentés dans les massifs paléozoïques français (notamment dans le Massif armoricain et en Montagne Noire au Sud du Massif central). Pourtant, la limite entre les deux systèmes n'est nulle part reconnue avec précision dans l'état actuel des investigations. Une lacune sédimentaire semble, en réalité, être assez généralisée, au moins pour la partie inférieure du Llandovery. Elle peut résulter de l'interférence d'un ensemble de causes, climatiques et variations eustatiques induites, épirogéniques et tectoniques distensives (échos taconiques) et manifestations volcaniques subordonnées.

Nous préciserons brièvement ces propos par l'examen de quelques successions.

Le Massif Armoricain

Différents domaines peuvent y être considérés.

En **Normandie,** la présence d'Ashgill est attestée par des Conodontes (zone à *Amorphognathus ordovicicus*) pour le Calcaire de Vaux (Weyant *et al.* 1977). Des fragments de ces calcaires sont repris dans la formation glacio-marine dite des 'pélites à fragments' ou Tillite de Feuguerolles qui est également rapportée à l'Ashgill supérieur grâce aux Chitinozoaires qu'elle renferme (F. Paris inédit). Dans les formations sus-jacentes l'absence apparente des Graptolites du Llandovery inférieur suggère une lacune correspondant au moins à celui-ci et débutant peut-être dans l'Hirnantien.

Dans **les parties centrales et orientales du Synclinorium médian armoricain,** la limite Ordovicien–Silurien se place entre les Formations de Saint-Germain-sur-Ille et de la Lande Murée (Fig. 1). Le passage entre ces deux formations est exposé dans diverses coupes des synclinoria du Ménez-Bélair et de Laval.

La Formation de Saint-Germain-sur-Ille, dans sa totalité, appartient à l'Ordovicien supérieur. Elle est habituellement subdivisée en deux unités lithologiques: un Membre inférieur à dominante arénacée, puissant de 200 m environ, et un Membre supérieur, argileux, et nettement moins développé (quelques dizaines de mètres d'épaisseur).

Des interlits argileux noirs s'intercalent dans l'ensemble grésoquartziteux constituant le Membre inférieur. Déposé dans un environnement littoral, voire tidal, ces grès livrent localement une abondante faune, généralement rassemblée dans des lits d'accumulation. On y reconnaît notamment des Brachiopodes (*Drabovinella erratica*), des Trilobites (*Calymenella bayani*, Homalonotidae), des Bivalves, et surtout des Graptolites qui ont permis de dater une partie de ce Membre inférieur (Skevington & Paris 1975). Ces Graptolites, limités à quelques niveaux gréso-micacés, sont exclusivement représentés par des Diplograptidae (*Orthograptus truncatus truncatus*, *O. truncatus abbreviatus*, *O. truncatus pauperatus* ainsi que de rares spéci-

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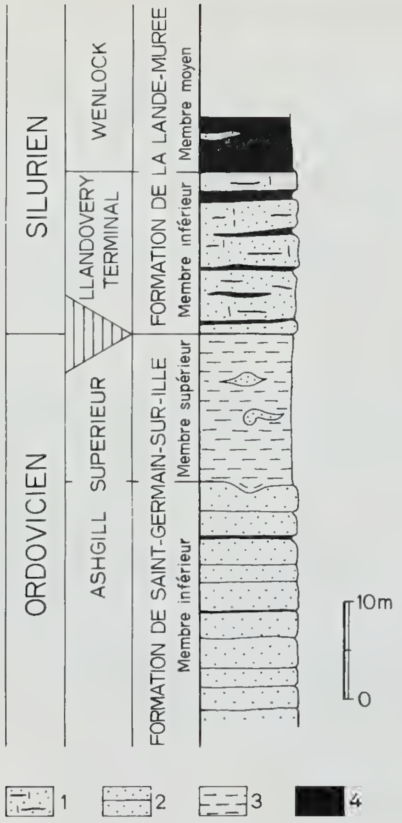


Fig. 1 Colonne stratigraphique à la limite Ordovicien–Silurien dans le Synclinorium du Menez-Belair. 1. Quartzites et mudstones. 2. Grès et quartzites. 3. Mudstones et siltstones. 4. Ampélites.

mens de ? *Climacograptus miserabilis* et de ? *Diplograptus fastigatus*). S'appuyant sur la fréquence relative des diverses sous-espèces de *O. truncatus*, Skevington & Paris (1975) admettent que les plus anciens niveaux à Graptolites de la Formation de Saint-Germain-sur-Ille appartiendraient à la partie supérieure de la Zone à *D. complanatus*, tandis que les niveaux les plus récents représenteraient la Zone à *D. anceps*. La partie supérieure du Membre inférieur de la formation a donc été attribuée à l'Ashgill. Les Trilobites étudiés par Henry (1980) et les Brachiopodes, revisés récemment par Mélou (1985), n'apportent pas de précisions stratigraphiques complémentaires. Quant aux Chitinozoaires, ils n'ont pas été observés dans les termes les plus élevés de ce Membre inférieur (Paris 1981).

Le Membre supérieur marque un net changement dans la lithologie. Sa base ravine le toit du Membre inférieur et ses caractères sédimentologiques (mudstones et siltstones noirs à 'ball and pillow structures') rappellent certains faciès des formations glacio-marines décrites dans l'Ordovicien terminal armoricain (Paris 1986). Aucune macrofaune n'y est connue. En revanche les Acritarches et les Chitinozoaires y sont relativement abondants. En dépit d'un état de conservation très médiocre, ces microfossiles évoquent des formes de l'Ashgill supérieur. Si l'on accepte un parallélisme entre ce Membre supérieur et des formations glacio-marines finiordoviciennes telles que la Formation des 'Pélites à fragments' de Normandie ou les argiles microconglomératiques du Nord de l'Afrique, le sommet de la Formation de Saint-Germain-sur-Ille appartiendrait à l'Ashgill supérieur et vraisemblablement à l'Hirnantien.

La Formation de la Lande Murée débute par un Membre inférieur constitué de quelques mètres de quartzites noirs, pyriteux, admettant des intercalations de mudstones à Graptolites, très riches en matière organique (ampélites) et montrant des teneurs anormalement élevées en éléments-traces (Dabard & Paris 1986). Le contact avec le Membre supérieur de la Formation de Saint-Germain-sur-Ille, correspondant à un brusque changement lithologique (Paris 1977), est plus ou moins bien exposé dans divers affleurements des synclinoria du Ménez-Bélair et de Laval (ex. carrière des 'Planches', en Guitté; carrière de 'Pont-Douve', en Médréac; carrière 'Pioc', en Vieux-Vy-sur-Couesnon; carrière du Rocher à Andouillé-Neuville; tranchée de l'autoroute Laval – Le Mans, à l'Ouest de Saint-Jean-sur-Erve; le 'Moulin du Feu' en Balazé).

Dans **le Synclinorium du Ménez-Bélair,** les premiers niveaux à Graptolites, parfois situés à moins d'un mètre au-dessus du contact entre les deux formations, appartiennent déjà au Telychien (sommet de la Zone à *turriculatus* ou Zone à *crispus*, selon les localités) (cf. Paris *et al.* 1980). Dans le Synclinorium de Laval, les premiers Graptolites récoltés dans la partie inférieure de la Formation de la Lande Murée appartiennent au Wenlock (Paris & Robardet, inédit). De toute évidence, il existe une lacune sédimentaire séparant les derniers dépôts ordoviciens (sommet du Membre supérieur de la Formation de Saint-Germain-sur-Ille) des premiers sédiments siluriens (base de la Formation de la Lande Murée). Cette lacune est d'ampleur variable selon les localités. Dans le Synclinorium du Ménez-Bélair, elle correspond au moins au Rhuddanien et à l'Aeronien (et peut-être au sommet de l'Ashgill). Dans le Synclinorium de Laval cette lacune paraît plus importante puisqu'elle implique l'ensemble du Llandovery et une partie du Wenlock.

Au Sud de Rennes, dans **le Synclinorium de Martigné-Ferchaud,** des travaux cartographiques (Herrouin, sous presse) ont récemment permis de préciser la succession lithologique locale, au voisinage de la limite Ordovicien–Silurien.

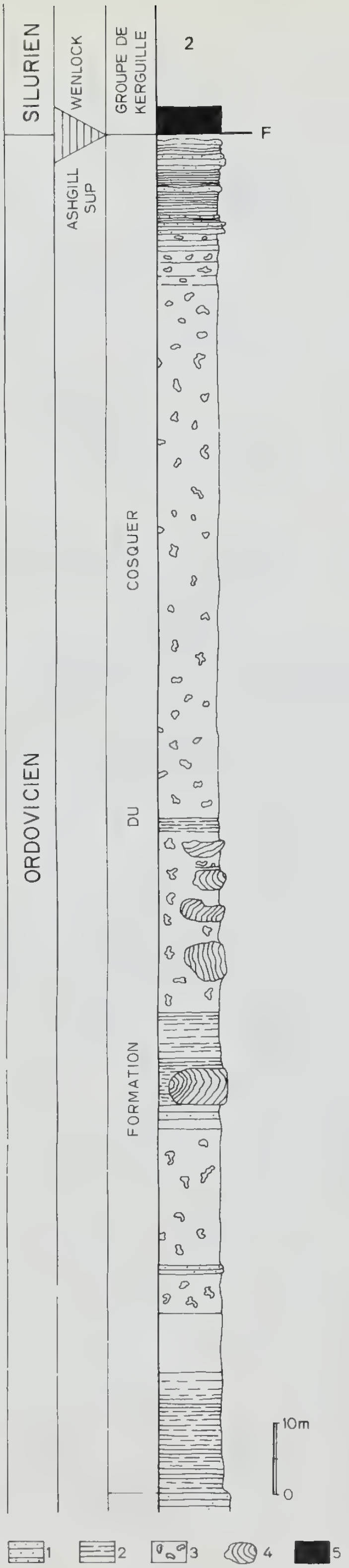
Succédant aux siltstones micacés, à lits gréseux, de la Formation de Riadan (traditionnellement rapportée au Caradoc et à l'Ashgill *pro parte*), on trouve la Formation de la Chesnaie (60 à 80 m de puissance). Cette unité comprend un ensemble inférieur gréso-quartziteux et une partie supérieure à dominante argileuse. Pour l'instant, la Formation de la Chesnaie n'a livré aucune faune exploitable. Au-dessus se placent les grès et quartzites blancs de la Formation de Poligné (60 à 70 m d'épaisseur). Le plus souvent azoïque, cet ensemble arénacé contient localement quelques Graptolites (Philippot 1950) de conservation trop médiocre pour fournir une attribution stratigraphique réellement fiable. Les premières faunes siluriennes significatives (Philippot 1950) apparaissent dans les mudstones noirs susjacents (ampélites). Il s'agit de riches assemblages de Graptolites de la base du Telychien (Zone à *turriculatus*).

Dans le Synclinorium de Martigné-Ferchaud, la limite Ordovicien–Silurien se place donc entre le toit de la Formation de Riadan et les ampélites de la base du Telychien. En l'absence de tout contrôle paléontologique rigoureux, la position de cette limite reste donc très approximative. Une lacune d'une partie du Llandovery, quoique vraisemblable, ne peut pour l'instant être démontrée.

Dans **la partie occidentale du Synclinorium médian,** la presqu'île de Crozon permet d'approcher la limite Ordovicien–Silurien dans deux contextes différents et tous deux incomplets.

La succession observée dans l'unité nord de la presqu'île (plage du Veryarc'h en Camaret) demeure d'interprétation difficile (Fig. 2). En concordance sur la Formation des Grès de Kermeur, datée du Caradoc dans sa partie moyenne (biozone 14 à *Jenkochitina tanvillensis*; Paris 1981), la Formation du Cosquer (Hamoumi 1981; Guillocheau 1983) débute par des shales noirs à lamines gréseuses, bien stratifiés, puis se caractérise par un ensemble à blocs glissés qui passent progressivement à des boules ('ball and pillow structures'). Les quartz de cette formation ont une origine glaciaire (Hamoumi *et al.* 1981). Vers le sommet, glissements et déformations s'atténuent, ce qui assure le passage à une stratification normale de grès à minces interlits de schistes noirs (Grès de Lamm-Saoz puissants de 6 mètres environ). Ces grès sont surmontés par les ampélites de la base du Groupe de Kerguillé qui livrent des Graptolites du Wenlock (Philippot 1950). La Formation du Cosquer n'a fourni aucun fossile dans les sédiments autochtones et son âge demeure imprécis. Un âge ashgillien a cependant été proposé (Paris *et al.* 1981) par comparaison notamment avec celui attribué à la formation glacio-marine de la Tillite de Feuguerolles de Normandie.

Les Grès de Lamm-Saoz furent, pour des raisons de géométrie, rapportés au Valentien (Silurien inférieur) par Philippot (1950). La récente découverte par l'un de nous (F.P.) de *Armor-*



SILURIEN
ORDOVICIEN
WENLOCK
ASHGILL SUP
GROUPE DE KERGUILLE
FORMATION DU COSQUER
2
F
10m
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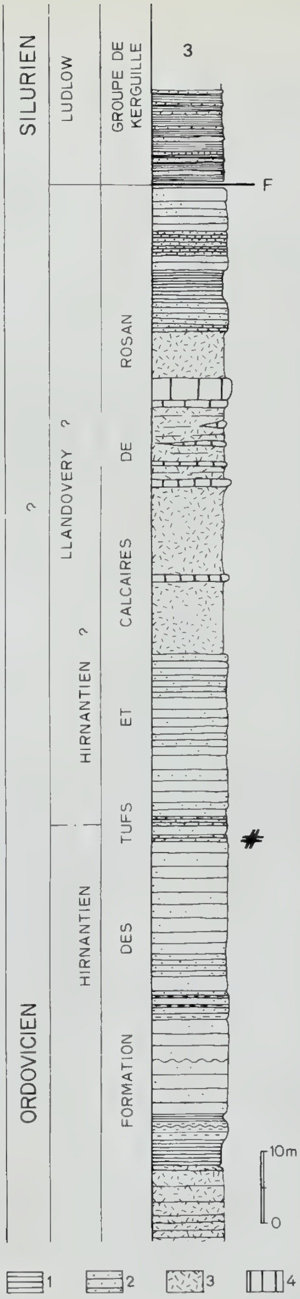

SILURIEN
ORDOVICIEN
LUDLOW
LLANDOVERY ?
HIRNANTIEN ?
?
HIRNANTIEN
GROUPE DE KERGUILLE
FORMATION DES TUFS ET CALCAIRES DE ROSAN
3
F
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Fig. 2 Contact Ordovicien–Silurien dans la coupe du Veryarc'h (Camaret, presqu'île de Crozon). 1. Grès et quartzites. 2. Shales. 3. Shales à 'ball and pillow structures'. 4. Megaslumps. 5. Ampélites.

Fig. 3 Contact Ordovicien-Silurien le long de l'Aulne, à l'Est de Tregarvan (presqu'île de Crozon). 1. Schistes. 2. Grès et quartzites. 3. Hyaloclastites. 4. Calcaires. # niveau à *Hirnantia*.

ichitina nigerica dans le dernier interlit noir de ces grès, situé à 30 cm sous les ampélites wenlockiennes, permet désormais de proposer, par comparaison avec les pélites à fragments du Sahara, un âge ashgillien supérieur pour la partie sommitale des Grès de Lamm-Saoz. Ainsi se trouve confirmée l'importance de la lacune qui correspond, dans cette unité nord, à la totalité du Llandovery.

Dans l'unité sud de la presqu'île de Crozon, la Formation des Grès de Kermeur est surmontée par un ensemble volcano-sédimentaire désigné Formation des Tufs et calcaires de Rosan. Aucun affleurement ne permet l'observation continue de la colonne correspondante. La base est concordante sur la Formation de Kermeur (falaise de Raguenez). Les coupes de la carrière du four à chaux de Rosan et de la route contiguë livrent en abondance *Nicolella actoniae*. La récente révision de cette espèce (Harper 1984) permet de considérer que nous sommes ici en présence de *N. actoniae ramosa*, sous-espèce de l'Ashgill. Ailleurs, à Lostmarc'h, les calcaires de Rosan sont également attribuables à l'Ashgill d'après les assemblages de Conodontes (zone à *Amorphognathus ordovicicus*) selon Paris *et al.* (1981). Un affleurement isolé le long de l'Aulne, à Coat-Garrec, a livré des Echinodermes (Chauvel & Le Menn 1972) qui ont confirmé l'âge ashgillien proposé pour cet affleurement par Mélou (1971) d'après la faune de *Leptestiina*. Enfin, il semble que la partie la plus élevée de cette formation soit représentée à l'Est de Trégarvan, le long de la rivière Aulne. La sédimentation carbonatée y régresse au profit des dépôts arénacés (Fig. 3). L'un de nous a récemment découvert dans cette coupe (Mélou 1987) un niveau à *Hirnantia sagittifera* au-dessus duquel 90 mètres de grès et de hyaloclastites avec quelques bancs carbonatés n'ont jusqu'à présent fourni aucun fossile. Cette partie sommitale de la formation peut donc encore correspondre à l'Hirnantien ou représenter déjà la base du Llandovery. La pile est tronquée par une faille importante qui la met en contact avec une partie élevée (Ludlow probablement) du Groupe de Kerguillé. Notons que ces observations nouvelles en presqu'île de Crozon tendent à réhabiliter un certain synchronisme des Formations du Cosquer et de Rosan qui avait été mis en doute récemment dans divers schémas (Paris *et al.* 1981; Guillocheau 1983).

Dans le Sud-Ouest du Massif armoricain, les données relatives à **l'unité vendéenne** demeurent fragmentaires (Ters 1979). Les schistes et grès schisteux rapportés à l'Ordovicien supérieur comme les schistes et phtanites à Radiolaires attribués au Llandovery n'ont pas livré de fossiles déterminants.

En conclusion, la présence de l'Ashgill, longtemps méconnue dans le Massif armoricain, y est désormais attestée dans plusieurs domaines et son extension inclut l'Hirnantien. Le Silurien, par contre, paraît en général amputé de sa partie basale au niveau d'une lacune qui peut, suivant les régions, intéresser Rhuddanien et Aeronien (Synclinorium de Martigné-Ferchaud, Synclinorium du Menez Bélair) ou affecter l'ensemble du Llandovery (presqu'île de Crozon). Le Massif armoricain ne permet donc aucune observation de la limite Ordovicien–Silurien.

Le Sud-ouest de la France

En **Aquitaine,** l'étude récente de sondages dans le socle paléozoïque sous la couverture mésocénozoïque, a permis à l'un de nous (F.P.) de constater, d'après les Chitinozoaires, la présence d'Ashgill terminal directement surmonté par des niveaux assez élevés du Llandovery. Une lacune du Silurien basal paraît donc également reconnaissable dans cette région.

En **Montagne Noire,** la succession de l'Ordovicien et du Silurien est observable en deux endroits connus depuis Chaubet (1937): au-dessus de la 'Tranchée noire' près de la Grange du Pin et au Petit Glauzy. De façon générale, la succession ordovicienne se termine par des alternances calcaréo-argileuses dites 'calcaires à Cystoïdes' et réputées d'âge ashgillien depuis Dreyfus (1948). Dans une récente révision des Brachiopodes de ces niveaux, Havlíček (1981) remet en cause cet âge et estime que les associations décrites indiqueraient plutôt le Caradoc supérieur. L'âge ashgillien demeure néanmoins plausible et si la faune à *Hirnantia* n'a pas été reconnue, les calcaires, quoique très pauvres en Conodontes, ont livré à l'un de nous (R.F.) quelques restes d'*Amorphognathus ordovicicus*. Ces niveaux terminaux, assez détritiques, n'ont fourni aucun Graptolite. Ceux-ci n'apparaissent que quelques mètres plus haut dans les argilites carburées. Dans la partie basale de ces schistes noirs, à la Grange du Pin, des Conodontes, extraits des nodules calcaires, indiquent selon Centène & Sentou (1975), la zone à *celloni* (équivalente des zones 20 à 23 des Graptolites, Llandovery moyen). Les mêmes niveaux livrent au Petit-Glauzy, selon ces auteurs, *Monograptus sedgwickii*, *M. uncinatus*, *M. nudus* du Llandovery moyen également (zone 21). On constate ainsi que la limite Ordovicien–Silurien ne peut être reconnue avec précision en Montagne Noire dans l'état actuel de la documentation. Faute de fossiles dans les niveaux qui assurent le passage entre les derniers carbonates à Cystoïdes et les premières ampélites à septaria, il demeure impossible de conclure à la continuité ou à l'existence de lacunes.

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The Ordovician–Silurian boundary in the Oslo region, Norway

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Synopsis

The Ordovician–Silurian boundary is exposed sporadically throughout the southern and central parts of the Oslo region; to the north there is an unconformity. In the central Oslo–Asker districts a well-developed *Hirnantia* fauna underlies beds with *acuminatus* Zone graptolites; other beds yield *Holorhynchus* faunas in the late Ordovician and early members of the *Stricklandia* lineage in the overlying Llandovery. Some early Silurian conodonts and acritarchs are recorded.

Lower Palaeozoic rocks outcrop in the Oslo region within a 230 km by 50 km area which is separated from the Precambrian to the east by the faults of a Permian graben. Within this broad region, most recent work in the late Ordovician and early Silurian has been achieved in the Oslo–Asker district, which lies in the approximate centre of the region, and also in the Hadeland district, some 50 km to the north of Oslo. These and other districts will be reviewed in turn. The Ordovician and Silurian beds in the area have been known since the early work of Murchison, Kjerulf, Broegger and others, and were the subject of a monumental study near the turn of the century by Kjaer (e.g. 1908). During the past fifteen years much new work has been done, for example Worsley *et al.* (1983) proposed a modern system of stratigraphical nomenclature for the Silurian rocks of the region.

Oslo–Asker District. The formation names for the late Ordovician stratigraphy (Fig. 1) were erected by Brenchley & Newall (1975), and its biostratigraphy and ecology elucidated by Brenchley & Cocks (1982), its trilobites described by Owen (1980, 1981) and its brachiopods by Cocks (1982). The topmost few metres of the Husbergøya Shale carries the trilobite *Tretaspis sortita broeggeri*, which Owen (1980) regarded as indicative of the uppermost Rawtheyan Stage. A *Hirnantia* fauna is known from horizons near the base of the Langøyene Sandstone Formation and within the Langåra Limestone–Shale Formation (Brenchley & Cocks 1982: 796), and includes common *Dalmanella testudinaria*, *Hirnantia sagittifera*, *Cliftonia* aff. *psittacina*, *Hindella cassidea*, *Eostropheodonta hirnantensis*, *Mucronaspis mucronata kjaeri*, bryozoans and criccoconariids, and less common *Acanthocrania*, *Glyptorthis*, *Lingula*, *Leptaena*, *Orbiculoidea*, *Oxoplecia*, *Philhedra*, *Calyptaulax*, *Illaenus*, *Platycoryphe*, *Toxochasmops*, molluscs, crinoids and carpoids. Elements of the *Hirnantia* fauna persist above this horizon in *Hindella–Cliftonia* and *Dalmanella* associations higher in the Langøyene Sandstone and there are also other faunas there such as one dominated by *Trematis norvegica* and modiolopsid bivalves. Above these, in the west of the area in Asker there occur thick beds largely composed of *Holorhynchus giganteus*, but with 13 other brachiopods and 17 other animals also recorded from them (Brenchley & Cocks 1982: 802), whilst in the east of the area, in the Oslo District, only trace fossils occur in rocks believed to represent a shore-face environment. At the top of the Langøyene Sandstone there occur shallow-water channel sequences which in some cases bear high abundance, low diversity faunas dominated by brachiopods such as *Brevilamnulella kjerulfi* and *Thebesia scopulosa*. This total sequence represents a regression since at least mid-Rawtheyan times, but above the channel-fill beds there occurs a metre-thick couplet of sandstone and limestone over the whole district which carries faunas, which are not age-diagnostic, of small shells such as *Onniella*, *Eoplectodonta*, *Leangella*, *Paucicrura* and *Dolerorthis*, as well as crinoids and bryozoa (and 16 other rarer forms). This couplet is lithologically included within the Langøyene Formation, but in fact marks the start of the 'early Silurian' transgression in the area. It is conform-

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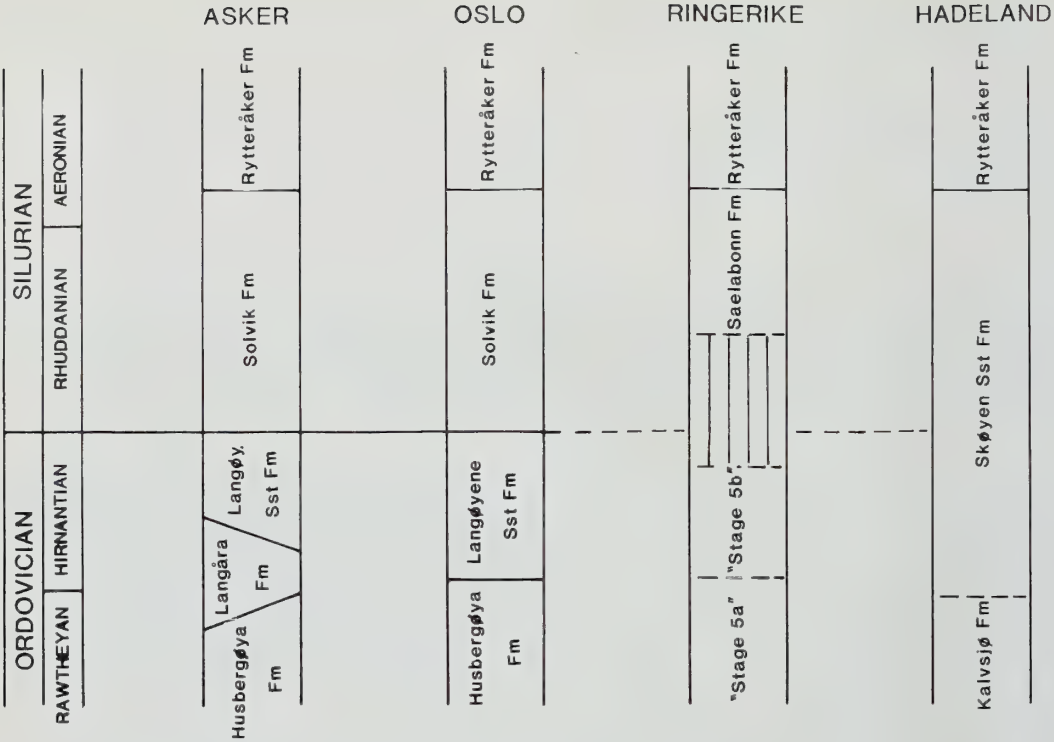

Fig. 1 Latest Ordovician and early Silurian stratigraphy in the Asker, Oslo, Ringerike and Hadeland districts of the Oslo Region.

ably followed by the basal organic-rich shales of the Solvik Formation in the Oslo District, which carries no shelly fauna, but from which Howe (1982) has identified *Climacograptus transgrediens* Waern from an horizon 11 m above the base of the formation at Ormøya, which he attributes to an horizon low in the *acuminatus* Zone (or perhaps high in the *persculptus* Zone). In the west of the area, in the Asker District, there was no break in the deposition of shelly faunas, and brachiopods are recorded from all three members of the Solvik Formation there, in a similar way to the higher parts of the formation in the Oslo District (Baarli 1985; Baarli & Harper 1986). The first occurrence of *Stricklandia lens prima* is at 95 m above the base of the Solvik Formation (Myren Member) and the transition from *S. lens prima* to *S. lens lens* occurs between 122 and 130 m above the base (Baarli 1986). Conodonts of the *Icriodella discreta–I. deflecta* Zone are known from 8 m above the base of the Solvik Formation at Konglungen, Asker (Aldridge & Mohamed 1982). Above the three members of the Solvik Formation, the Rytteråker Formation yields pentamerides and conodonts of Aeronian age: Nakrem (1986) has identified the *Distomodus kentuckyensis–D. staurognathoides* conodont zonal boundary as occurring at about the boundary of the Solvik and Rytteråker Formations.

The Ringerike area. The latest Ordovician of the Ringerike area remains unrevised, and thus the old stage terminology of Kiaer (1897, 1908) is employed—it carries a rich brachiopod fauna, but one not identical to that from the Oslo–Asker region and no *Hirnantia* fauna is known from the area; it also differs in the presence of bioherms and patch reefs within Stage 5b, the most notable of which is at Ullerntangen. The relationships between the Ordovician and Silurian beds are obscure and a local unconformity is postulated here (Fig. 1). The overlying beds of the Saelabonn Formation are shallow-water storm deposits with lenses of displaced shelly faunas (Thomsen 1982); their detailed age is indeterminate, but probably includes the Lower Llandovery. The overlying Rytteråker Formation includes the *Borealis–Pentamerus* transition near its base (Mørk 1981), and that horizon is certainly now in the Aeronian. Smelrør (1987) has identified the acritarch zones 1 and 2 of Hill (1974) as occurring in the Saelabonn Formation.

The Hadeland area. Owen (1978) has revised the late Ordovician and early Silurian of this area and established a Rawtheyan age for the Kjørrven Formation which underlies the Kalvsjø Formation, which carries a sparse trilobite fauna, some brachiopods and the cystoid *Hemicosmites* and the calcareous alga *Palaeoporella* which indicate an Ordovician rather than a Silurian age. Above this the 120 m thick Skøyen Sandstone Formation appears to straddle the Ordovician–Silurian boundary, since beds with *Zygospiraella* and other typical early Silurian brachiopods occur from about the middle of the formation. The Skøyen Sandstone is succeeded conformably by the Rytteråker Formation which yields *Borealis borealis* near its base.

Other areas. From the Skien and Porsgrunn district near the south of the Oslo Region, for example in a section at Herøyavegen, Porsgrunn, *Holorhynchus* beds followed by early Silurian beds yielding *Zygospiraella duboisi* (Verneuil) and *Eostropheodonta mullochensis* (Reed) are known, but the stratigraphy is unrevised. In the Oslo region north of Hadeland there is an unconformity between the late Caradoc and early Ashgill Mjøesa Limestone and the early Silurian, for example Møller (1986) has described the succession at Brummunddal, where the Helgøya Quartzite of probable Aeronian age bearing *Borealis borealis* rests on the Mjøesa Limestone.

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East Baltic Region

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Synopsis

Five confacies belts from north to south, from Estonia through Latvia to Lithuania, are described briefly through the late Ordovician and early Silurian, with their varied facies and faunas. Despite clear breaks corresponding to the Ordovician–Silurian boundary at the edges of the depositional basin, rocks of Hirnantian age are identified from the centre of the basin, including *Hirnantia* and *Dalmanitina* faunas in the Porkuni Regional Stage and basal Silurian faunas, including some graptolites, chitinozoans, brachiopods and conodonts, from the overlying Juuru Regional Stage. Any stratigraphical break at the boundary appears to be represented by no more than a facies change.

Introduction

The East Baltic area is a part of the extensive gulf-like Baltic sedimentary basin (Männil 1966; Kaljo & Jürgenson 1977). The uppermost Ordovician and the lowermost Silurian are mostly represented by carbonate or terrigenous-carbonate rocks with an exceptionally rich benthic shelly fauna; however, pelagic groups of fossils, especially graptolites, are of a more restricted distribution. The rocks are tectonically undisturbed, and unmetamorphozed (CAI 1–1·5), with only a little dolomitization in places, and the fossils are well preserved. The bedding is almost horizontal and dips slightly to the centre of the basin. The distribution of the Ashgill–Llandovery rocks in the East Baltic is shown in Fig. 1. The outer margin of the area is erosional and corresponds to the base of the Ashgill (Vormsi Regional Stage). The axial part of the basin with the most deep-water rocks corresponds to the Baltic Syneclise (IV belt), and along its margins there occur shallower-water sediments.

Most of the area is covered by younger rocks. The outcrops of the Ordovician–Silurian boundary strata are confined to North Estonia (Belt I in Fig. 1), where only comparatively shallow-water deposits are exposed. A more complete succession of facies in the basin can be seen in borehole sections. Fig. 2 presents a cross section of Ashgill and Lower and Middle Llandovery strata along the Orjaku–Remte–Ukmerge line, which is shown in Fig. 1. The section goes across the main facies belts of the basin and shows the relations between local lithostratigraphical units and their general lithology. In the figure stratigraphical units are marked with letter-indexes: their full nomenclature is given in Fig. 3.

Confacies belts

In the East Baltic five confacies belts can be distinguished in Ordovician–Silurian boundary beds. Their distribution is shown in Fig. 1 and their lithological composition in Fig. 2.

Type I—the most shallow-water sections in North Estonia and Hiiumaa Island represented by aphanitic, bioclastic and biohermal limestones. In the Raikküla Formation there occur primary argillaceous dolomites of lagoonal origin in places. Some considerable stratigraphical gaps have been established (Fig. 3). The Ordovician ends with Early Porkuni bioclastic, biohermal and arenaceous limestones (Ärina Formation), which carry a *Streptis* brachiopod community (Hints 1986), disconformably overlain by Juuru aphanitic (Koigi Member) and biomicritic limestones (Varbola Formation) with a *Stricklandia* community (Rubel 1970).

Type II—sections in central Estonia and Saaremaa Island. Represented by marls, aphanitic and biomicritic nodular limestones. The sections are more complete than in Type I. A distinct hiatus has been established only in the upper part of the Porkuni Regional Stage and in the west at the top of the Raikküla Stage. The Ordovician–Silurian boundary interval is similar to the sections of Type I, but southwards the Ärina Formation and the Koigi Member thin out

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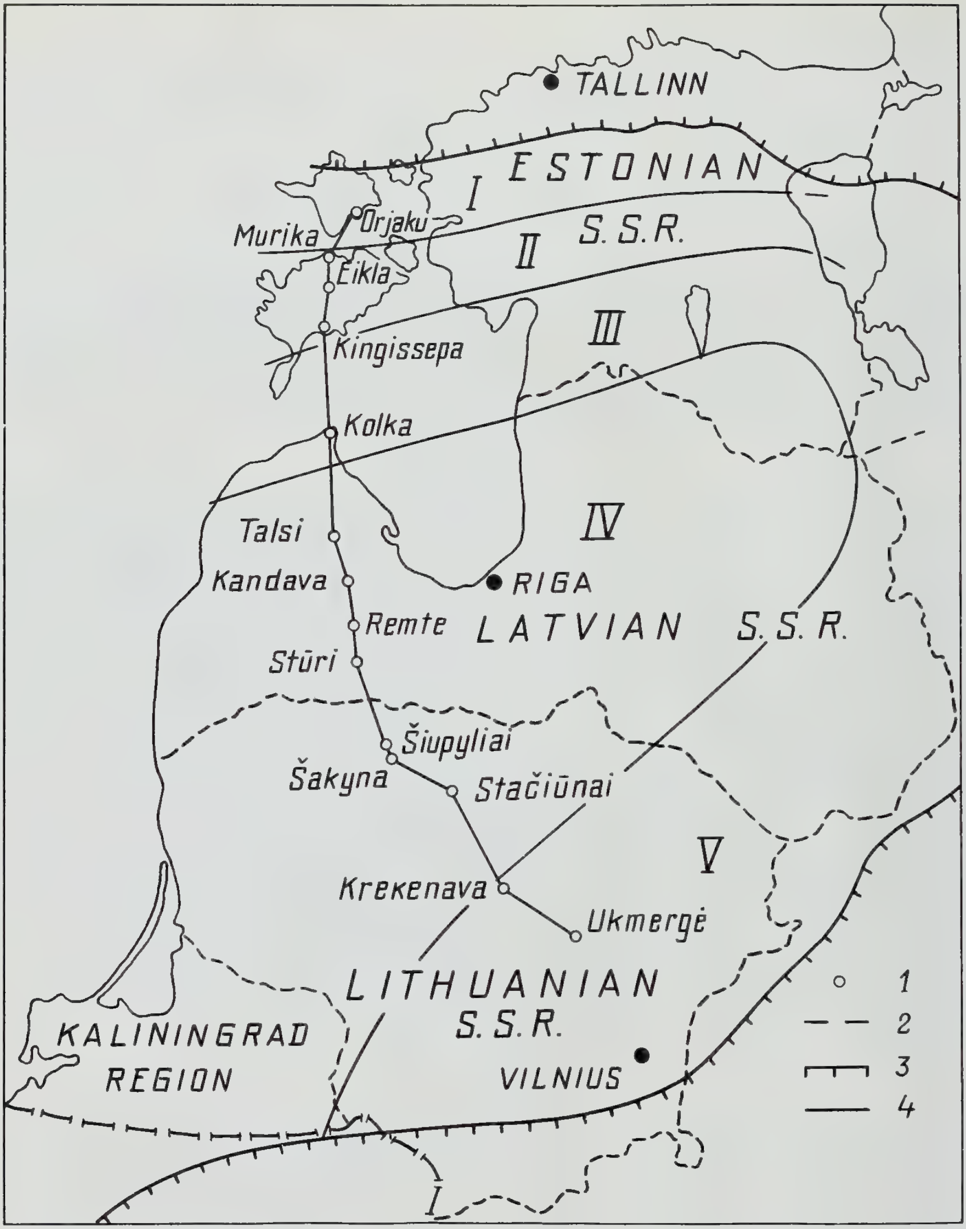
TALLINN
ESTONIAN
S.S.R.
I
II
III
IV
V
Murika
Orjaku
Eikla
Kingissepa
Kolka
Talsi
Kandava
RIGA
Remte
LATVIAN S.S.R.
Stūri
Šiupyliai
Šakyna
Stačiūnai
Krekenava
Ukmergė
LITHUANIAN S.S.R.
KALININGRAD REGION
VILNIUS
1
2
3
4

Fig. 1 Distribution of Ordovician–Silurian boundary rocks in the East Baltic area. 1—boreholes, 2—administrative boundaries, 3—outer margin of the distribution of Ashgill and Llandovery rocks, 4—boundaries of main types of sections, marked with Roman numbers.

and the boundary of the systems continues in a comparatively monotonous complex of nodular limestones and marls. In places the Porkuni Regional Stage may be missing.

Type III—sections in south Estonia and north-west Latvia. Marls and argillaceous limestones, including their red-coloured varieties, are significant lithologies. In the Llandovery, aphanitic limestones alternate with marls. A considerable erosional gap corresponds to the upper part of the Pirgu Regional Stage, and this gap increases westwards. The uppermost

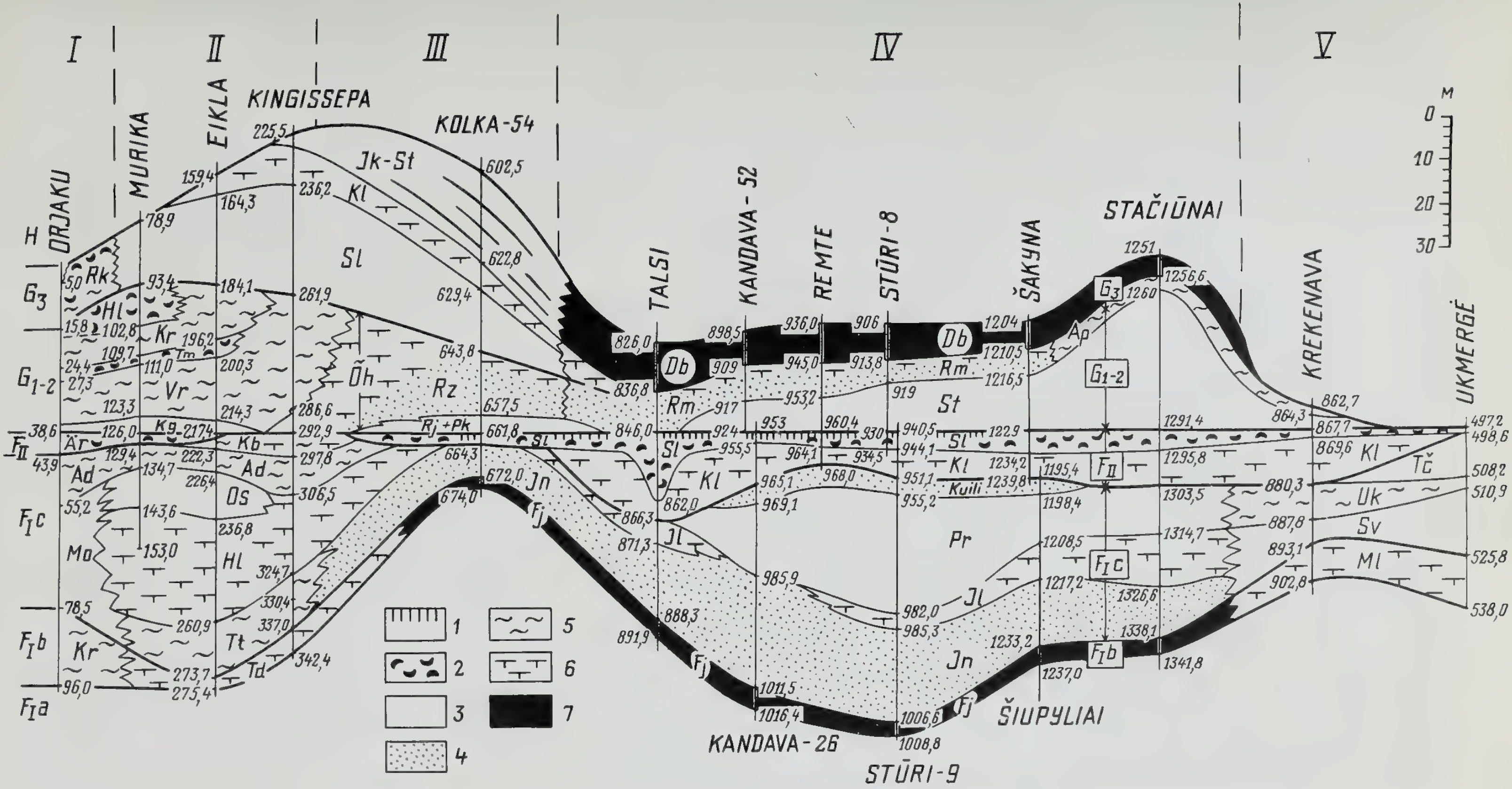
I
II
III
IV
V
ORJAKU
MURIKA
EIKLA
KINGISSEPA
KOLKA-54
TALSI
KANDAVA - 52
REMTE
STŪRI-8
ŠAKYNA
STAČIŪNAI
KREKENAVA
UKMERGĖ
KANDAVA - 26
STŪRI-9
ŠIUPYLIAI
H
G_3
G_{1-2}
F_{II}
F_Ic
F_Ib
F_Ia
Rk
Hl
Kr
Tm
Vr
Kg
Ar
Ad
Mo
Kb
Os
Hl
Tt
Td
Jk-St
Kl
Sl
Õh
Rz
Rj + Pk
Jn
Fj
Db
Rm
St
Kl
Kuili
Jl
Pr
Ap
Uk
Tč
Sv
Ml
1
2
3
4
5
6
7
M
0
10
20
30

Fig. 2 Stratigraphical cross section of the boundary interval along the line shown in Fig. 1. 1—thin-bedded argillaceous-carbonate rocks of lagoonal origin; 2—biosparitic, clastic and oolitic limestones (sparitic calcarenites); 3—fine- to cryptocrystalline limestones (calcilutites); 4—red argillaceous limestones and marls; 5—biomicritic nodular limestones; 6—grey argillaceous limestones and marls; 7—graptolitic mudstones (shales). For meaning of the stratigraphical indexes see Fig. 3: the index of a formation consists of the first letter and the following first consonant (if any) of the name.

Ordovician is represented by marls and argillaceous limestones with the *Dalmanitina* Fauna (Kuldiga Formation). Above this occur biosparitic, oolitic and arenaceous limestones of the Saldus Formation. The Silurian begins with marls and argillaceous limestones of the Õhne Formation with the *Clorinda* community (Rubel 1970).

Type IV—sections in southeast Estonia, considerable part of Latvia, west Lithuania and the Kaliningrad Region. The studied stratigraphical interval begins and ends with dark graptolitic mudstones with the assemblage of the *Pleurograptus linearis* Zone in the Ordovician part (Fjäcka Formation) and of the *Coronograptus cyphus–Monograptus sedgwickii* Zones in the Silurian (Dobele Formation). Between these key beds there occur red and grey calcareous mudstones, marls and aphanitic limestones. The uppermost Ordovician is analogous to the sections of Type III. The Silurian begins with marls and aphanitic limestones of the Stačiunai Formation which have yielded few fossils good for correlation.

Type V—sections in east Lithuania and southeast Latvia with an extensive hiatus at the boundary interval. More or less continuous Upper Ordovician deposits are represented by marls and various limestones which end at the top of the Pirgu Regional Stage with the aphanitic limestones of the Taučionys Formation which yield a *Holorhynchus* fauna. There is a hiatus at the level of the Porkuni, Juuru and Raikküla Regional Stages, or in places there occur thin residual tongues and lenses of the Kuldiga, Saldus and Apasčia Formations, which are transgressively overlain by mudstones and marls of the late Llandovery Adavere Regional Stage.

In the westernmost part of Lithuania and in the Kaliningrad District the rocks of the Ordovician–Silurian boundary interval become still more argillaceous and graptolites occur throughout the whole section, with the exception of the uppermost Ordovician which yields a shelly *Hirnantia* fauna. This is a transition to a different type of facies belt which is distributed in north Poland and the southern part of the present Baltic Sea.

Thus analysis of the lithologies and fossils of the various sections shows that by the end of the Ordovician the Baltic basin had experienced a considerable regression which reached its maximum in the second half of the Porkuni. This is indicated both by hiatuses in the sections (Fig. 3) and by the presence of calcareous oolites and early diagenetic (or sedimentary) dolomi-

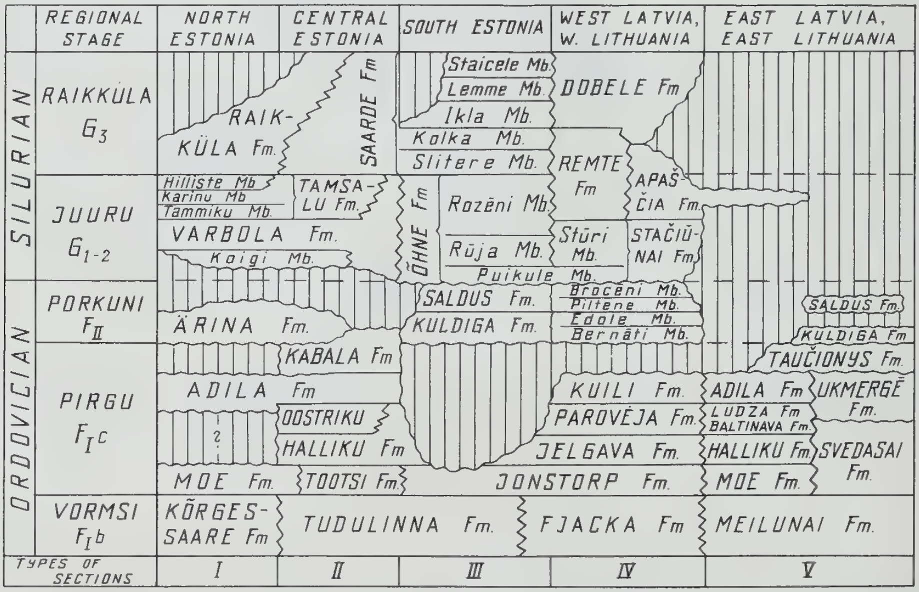
REGIONAL STAGE
NORTH ESTONIA
CENTRAL ESTONIA
SOUTH ESTONIA
WEST LATVIA, W. LITHUANIA
EAST LATVIA, EAST LITHUANIA
SILURIAN
ORDOVICIAN
RAIKKÜLA G_3
JUURU G_{1-2}
PORKUNI F_{II}
PIRGU F_Ic
VORMSI F_Ib
RAIK-KÜLA Fm.
SAARDE Fm
Staicele Mb.
Lemme Mb.
Ikla Mb.
Kolka Mb.
Slitere Mb.
DOBELE Fm
REMTE Fm
APAŠČIA Fm.
Hilliste Mb.
Karinu Mb.
Tammiku Mb.
TAMSA-LU Fm.
ÕHNE Fm
Rozēni Mb.
VARBOLA Fm.
Kaigi Mb.
Rūja Mb.
Stūri Mb.
STAČIŪ-NAI Fm.
Puikule Mb.
SALDUS Fm.
Broceni Mb.
Piltene Mb.
Edole Mb.
Bernāti Mb.
ÄRINA Fm.
KULDIGA Fm.
SALDUS Fm.
KULDIGA Fm
KABALA Fm
TAUČIONYS Fm.
ADILA Fm
KUILI Fm.
ADILA Fm
UKMERGĖ Fm.
?
OOSTRIKU
PARОVĖJA Fm.
LUDZA Fm
BALTINAVA Fm.
HALLIKU Fm
JELGAVA Fm.
HALLIKU Fm
SVEDASAI Fm.
MOE Fm.
TOOTSI Fm.
JONSTORP Fm.
MOE Fm.
KÕRGES-SAARE Fm
TUDULINNA Fm.
FJACKA Fm
MEILUNAI Fm.
TYPES OF SECTIONS
I
II
III
IV
V

Fig. 3 Stratigraphical scheme of the late Ordovician and early Silurian boundary rocks in the East Baltic area.

tes in the Saldus Formation in the axial part of the basin. The character of the transition from the Porkuni to the Juuru Regional Stage and the lithology of the sequences indicate a rapid deepening of the basin, obviously of glacial eustatic origin (Kaljo *et al.*, in press).

Local stratigraphy

Knowledge of the local stratigraphy of rocks near the Ordovician–Silurian boundary has considerably improved in the past few decades. The correlation chart presented in Fig. 3 is based on the decisions of the regional stratigraphical conferences in Vilnius in 1976 and in Tallinn in 1984 (Grigelis 1978). The chart was compiled from material in many publications (see further references in the papers by Männil 1966, Kaljo 1970, Kaljo & Klaaman 1982, Paškevičius 1979, Grigelis 1982, Ulst *et al.* 1982).

Dynamics of the faunas

From the five regional stages from Vormsi to Raikküla which correspond to the Ashgill and lower and middle Llandovery, extremely rich fossil faunas have been collected. The present paper uses the data obtained through the study of eight groups of fossils: stromatoporoids, tabulate corals, brachiopods, trilobites, ostracodes, chitinozoans, conodonts and graptolites. In total 734 species from 313 genera and 105 families have been identified. Table 1, which is based on data by Nestor *et al.* (in press), shows the distribution of species and genera by stages. It shows that the associations of the Porkuni and Juuru Regional Stages are the least diverse; and also that they have almost no common species, whereas about one third of the genera occur in both stages. At the Ordovician–Silurian boundary, besides intensive extinction of the Ordovician fauna, the rate of the appearance of new fauna also rose. In Porkuni times extinction prevailed and Juuru times were characterized by the appearance of new faunas.

Table 1 Numbers of species and genera of eight fossil groups recorded from the Vormsi to Raiküla Regional Stages.

Regional Stage	Vormsi	Pirgu	Porkuni	Juuru	Raikküla
Species, total number	195	252	154	177	221
transitional from the underlying stage, %	43	38	17	4	22
Genera, total number	150	175	125	109	130
transitional from the underlying stage, %	57	69	49	32	57

The dynamics of the fossil groups varied according to their ecology. For example, the shallowing of the basin in the Late Ordovician led to the radiation of the shallow-water stromatoporoids and corals, whereas the graptolites emigrated completely from the East Baltic area at the same time as the general crisis of graptolites noted by Rickards (1978) became apparent. Shallowing was also of great influence on the benthic trilobites and ostracodes, which usually inhabited deeper shelf areas and a remarkable decrease in their diversity took place in Pirgu and Porkuni times. The reverse tendency can be seen during the rapid deepening of the basin at the beginning of Juuru times; however, at that time shallow-water groups, particularly stromatoporoids and corals, were chiefly affected.

Biostratigraphy and correlation

Space does not allow a more detailed analysis here of the diverse biota from the boundary beds, and so only selected lists of species for each stage are presented, those which are most valuable for correlation (in brackets the index of the formation is shown where the species has been found).

Vormsi Regional Stage
Catenipora wrighti Klaamann (Kr), *Plaesiomys solaris* Buch (Kr), *Kullervo complectens* (Wiman) (Td), *Acanthochitina barbata* Eisenack (Td, Fj, Ml), *Tretaspis seticornis* (Hisinger) (Fj), *Orthograptus quadrimucronatus* (Hall) (Fj), *Climacograptus styloides* Lapworth (Fj), *Hamarodus estonicus* Viira (Fj), *Belodina compressa* (Branson & Mehl) (M1).

The above species enable a clear determination of the position of the Stage at the level of the graptolite *Pleurograptus linearis* Zone.

Pirgu Regional Stage
In the lower part: *Eospirigerina sulevi* (Alichova) (Mo, Jn, Sv), *Foramenella parkis* (Neckaja) (Mo, Jn, Sv, Ad, Uk), *Amorphognathus ordovicicus* Branson & Mehl (Mo, Jl), *Dicellograptus* cf. *complanatus* Lapworth (Mo), *Rectograptus gracilis* (Roemer) (Hl, Jn), *Panderia megalophthalma* (Linnarsson) (Jn), *Tretaspis latilimba* (Linnarsson) (Jn, Jl, Kl).

In the middle part: *Clathrodictyon microundulatum* Nestor (Ad), *Catenipora tapaensis* (Sokolov) (Ad), *Esthonia asterisca* Roemer (Ad, Uk), *Maclurites neritoides* (Eichwald) (Ad), *Belodina compressa* (Branson & Mehl) (Ad).

In the topmost part: *Conochitina taugourdeaui* Eisenack (Kb), *Climacograptus supernus* Elles & Wood (Kb), *Holorhynchus giganteus* Kiaer (Tč).

The graptolites shown above enable a correlation of the stage with the zones of *Dicellograptus complanatus* and *D. anceps.*

Porkuni Regional Stage
Paleofavosites rugosus Sokolov (Är), *Rhabdotetradium frutex* Klaamann (Är), *Streptis undifera* (Schmidt) (Är), *Illaenus angustifrons depressa* Holm (Är), *Apatochilina falocata* Sarv (Är), *Dalmanella testudinaria* (Dalman) (Kl), *Hirnantia sagittifera* (M'Coy) (Kl), *Eostropheodonta hirnantensis* (M'Coy) (Kl), *Dalmanitina* (*Mucronaspis*) *mucronata* (Brongniart) (Kl, Sl), *Brongniartella platynota* (Dalman) (Kl), *Pseudulrichia norvegica* Henningsmoen (Kl), *Conochitina postrobusta* subsp. A (Nõlvak, Ms).

The representatives of the *Hirnantia* and *Dalmanitina* communities enable correlation with the Hirnantian Stage at the level of the zones of *Climacograptus extraordinarius* and *Glyptograptus persculptus.*

Juuru Regional Stage
Clathrodictyon boreale Riabinin (Vr, Tm), *Paleofavosites paulus* Sokolov (Vr, Tm, Õh), *Stricklandia lens prima* Williams (Vr, lower pt), *S. lens lens* Williams (Vr, upper pt), *Borealis borealis* (Eichwald) (Tm), *Calymene ansensis* Männil (Vr, Tm), *Acernaspis estonica* Männil (Õh), *Aitilia senecta* Sarv (Vr), *Steusloffina eris* Neckaja (Vr, Tm, Õh), *Ozarkodina* ex gr. *oldhamensis* (Rexroad) (Õh, lower pt), *Distomodus* cf. *kentuckyensis* Branson & Branson (Õh), *Ancyrochitina laevaensis* Nestor (Õh, lower pt), *Conochitina postrobusta* Nestor (Õh), *Dimorphograptus confertus* (Nicholson) (Õh, upper pt), *Pribylograptus incommodus* Törnquist (Õh, top).

The top of the Juuru Regional Stage is well defined by graptolites, suggesting that this level approximately coincides with the boundary of the *Dimorphograptus confertus* (equivalent to the *Orthograptus vesiculosus*) and *Coronograptus cyphus* Zones (Kaljo *et al.* 1984). The age of the lower limit of the stage can be established by *Stricklandia lens prima* (according to Cocks, 1971, it equates to the level of the *Parakidograptus acuminatus* Zone) and by the listed chitinozoans and conodonts, indicating that there was no substantial regional hiatus at the base of the Silurian in the East Baltic. However, distinct breaks occur at the margins of the basin, particularly to the southeast.

The correlation of the Raikküla Regional Stage is clearly defined by graptolites within the *Coronograptus cyphus* and *Demirastrites convolutus* Zones (Kaljo 1967; Kaljo 1970; Kaljo *et al.* 1984). Detailed correlations in Estonia were considerably improved by the study of chitinozoans (Nestor 1976).

The present data from graptolites and other evidence permit only general correlation of the East Baltic section with the Dob's Linn section, but finds of *Climacograptus supernus* at the top of the Pirgu and *D. confertus* at the top of the Juuru Regional Stage do not contradict the placing of the Ordovician–Silurian boundary (the base of the *P. acuminatus* Zone) at the top of the Porkuni Regional Stage.

Correlation with the Anticosti section is possible by means of chitinozoans and conodonts. In this section (Achab 1981; McCracken & Barnes 1981) Member 5 of the Ellis Bay Formation is characterized by the presence of *Conochitina taugourdeaui*, *C. micracantha* and *C. gamachiana*. J. Nõlvak has found the first two and a form similar to the third species at the top of the Pirgu Regional Stage. At the base of Member 6 in Anticosti *Ozarkodina oldhamensis* appears, and somewhat higher *Distomodus kentuckyensis* and above bioherms *Ancyrochitina spongiosa* are recorded. P. Männik, V. Nestor and V. Viira have found all these species or closely related forms in the lower part of the Juuru Regional Stage. Thus, in the Anticosti section we do not see equivalents of the Porkuni Regional Stage (at least of its upper part) which is characterized by *Conochitina postrobusta* subsp. A.

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The Ordovician–Silurian boundary in Poland

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Synopsis

Outcrops in the Holy Cross Mountains and Sudetes, as well as boreholes in the Polish lowlands, show Ordovician–Silurian boundary sediments to be variably developed or sometimes absent. The *Hirnantia* fauna is developed, but most other rocks are in graptolitic facies.

Ordovician–Silurian boundary beds have been recognized in Poland both in outcrops and in boreholes. However, despite abundant documentation obtained from both types of sections as well as intensive investigations carried out, the boundary in Poland is still inadequately known. This is mainly because of the presence of many sedimentary gaps in the known sections, which are a result of the Taconic orogenic phase, and also because of the lack of good index fossils. In consequence, this boundary is not sharply defined in the Polish profiles, which makes good correlation with the adjacent regions difficult (Teller 1969).

The Ordovician–Silurian boundary beds outcrop in Poland only in the Holy Cross Mountains and in the Sudetes. In the Bardo Range of the Sudetes (Teller 1962) there are no fossils known from near the junction, so the boundary has been arbitrarily designated by the presence of Lower Llandovery graptolites in black siliceous shale among the liddites The upper Ordovician sediments appear to be represented in this area by alternating beds of sandstone and shale without fossils which underlie the Silurian liddites. The Ordovician–Silurian boundary has been put at the contact of these two formations, but it is not known for certain whether or not the clastic Ordovician corresponds to the uppermost Ashgill.

In the Holy Cross Mountains, the boundary beds are known to occur in the Zalesie profile (Kielan 1956, 1957; Temple 1965), in the southern limb of the Bardo syncline in the Kielce region. The uppermost Ashgill silty beds contain a *Hirnantia* fauna with *Mucronaspis mucronata* Brongniart), *M. olini* Temple, *Dalmanella testudinaria* (Dalman), *Hirnantia sagittifera* (M'Coy) and *Eostropheodonta hirnantensis* (M'Coy) amongst others, and are covered by black shales with *Akidograptus acuminatus* at their base, accompanied by *Climacograptus scalaris normalis* and *A. ascensus*, indicating the *acuminatus* Zone.

Thus the boundary separates the Upper Ashgill siltstone formation, containing a *Hirnantia* fauna, from the Lower Llandovery black shale formation with graptolites. This rapid change in facies suggests a lack of sedimentary continuity particularly since there are no graptolites in the uppermost Ashgill. In profiles in other parts of the world, the *Hirnautia* fauna (Cocks 1985) is generally older, or is to be found below the Ordovician *Glyptograptus persculptus* Biozone, the top of which is now taken as the boundary between the Ordovician and the Silurian.

In many other sections in the Holy Cross Mountains (Tomczyk 1962; Bednarczyk 1973) a sedimentary gap is noted at this boundary. This gap embraces the entire Upper and partly the top of the Lower Ashgill as well as the lowermost Llandovery, and appears to be a result of the Taconic phase of orogeny.

In the Polish Lowlands, the Ordovician–Silurian boundary beds show great facies variability (Modliński 1973). In many boreholes, sedimentary gaps embrace various time spans and a change of facies toward a marly-arenaceous one is noted, which appears to indicate a gradual regression. Graptolites have only been found in the deeper parts of the platform slope clayey facies, including the Upper Ashgill Biozone of *Glyptograptus persculptus* and the Lower Llandovery *A. acuminatus* Zone, for example in the Lebork borehole (Tomczyk 1965).

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The Ordovician–Silurian boundary in the Prague Basin, Bohemia

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Synopsis

The Ordovician–Silurian boundary in the Prague Basin is marked by an abrupt change in facies development and faunal assemblages, without significant breaks in purely marine sedimentation. Shallow marine sandstones and petromict conglomerates of the upper Kosov (Hirnantian) are followed by bioturbated mudstones due to the initial phase of a new transgression, with an abundant *Hirnantia* fauna in the uppermost Kosov. The mudstones are followed by dark graptolitic shales at the base of the Silurian (in the Prague Basin at the base of the *Akidograptus ascensus* Subzone). During the *Parakidograptus acuminatus* Subzone another change of sedimentation appeared as a transition from silty-clay shales to sandy-micaceous laminites. This change corresponds to a local break in sedimentation in the north limb of the Prague Basin and in the Pankrác area, where the break continued to the *Monoclimacis griestoniensis* Zone. The sequence and the succession of faunal assemblages indicate an accelerated rate of transgression just below and above the Ordovician–Silurian boundary. Analysis of the faunal assemblages allows a detailed stratigraphical subdivision of the boundary beds in the Prague Basin and wide international correlation.

Introduction

In Bohemia, the Ordovician–Silurian boundary is well developed in the Prague Basin (Barrandian area). The Prague Basin is a tectonically predisposed linear sedimentary depression in which the sedimentation continued from the lowermost Ordovician up to the Middle Devonian without substantial interruptions (Havlíček 1981, 1982). In the Prague Basin, the Ordovician–Silurian boundary coincides with the boundary between the Kosov and Želkovice Formations. Perner & Kodym (1919) supposed that there was a stratigraphical gap at the base of the Silurian in Bohemia caused by the emersion phase of the Taconic orogeny. Later, the lowermost Silurian graptolite zones, including the *Parakidograptus acuminatus* Zone, were documented in the Barrandian area by Marek (1951) and by Bouček (1953) in an isolated outcrop near Běchovice. These authors denied the existance of the boundary gap east of Prague at Běchovice, but they admitted its presence in the rest of the Prague Basin. Horný (1956, 1960) found the earlier *A. ascensus* Zone along the whole southern limb of the Basin. He recorded that the rocks of the basal Silurian graptolite zones were only absent locally due to minor erosion caused by epeirogenetic movements that represented the aftermath of tectonic activity during the deposition of the Kosov Formation. Havlíček (1981, 1982) explained both the flysch-like Kosov Formation and the change in lithologic development at the Ordovician–Silurian boundary by invoking synsedimentary tectonic movements in the basin.

More recently, basal Silurian graptolite zones have also been discovered in the northern limb of the Prague Basin and the boundary hiatus was verified only in a restricted part of the basin (Štorch 1982, 1986). Investigation of the early Kosov (Štorch & Mergl, in press) has shown the sequence in Bohemia to be very similar to that explained by glacio-eustatic environmental changes (Brenchley & Cocks 1982; Brenchley & Cullen 1984; Brenchley & Newall 1984). The glacio-eustatic conception of the late Ordovician to early Silurian facies and faunal changes (Brenchley 1984; Brenchley & Newall 1984) also appears to explain the Ordovician–Silurian boundary sequence in the Prague Basin.

Sequence of the latest Ordovician

Considerable changes preceding the Ordovician–Silurian boundary event were recorded at the top of the Králův Dvůr Series in the Prague Basin (Štorch & Mergl, in press). The deep water

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mudstones of the Králův Dvůr Formation, with deep water faunal assemblages, were followed by coarse grained subgraywackes and silty shales at the base of the Kosov Formation. The high-diversity *Proboscisambon* Community of the uppermost Králův Dvůr Formation was replaced by the low-diversity and short-lived *Mucronaspis* Community (Štorch & Mergl, in press), the last record of which (bivalves and trilobite fragments) occurs in the shale of the lowermost Kosov Formation.

The basal Kosov subgraywackes and shales were succeeded by flysch-like sediments which form most of the thickness of the Kosov Formation. This regressive sequence culminated in the deposition of shallow-water sandstones and petromict conglomerates in the upper part of the formation. In the uppermost sandstone layers a monotonous assemblage of infaunal bivalves provides evidence of intertidal environments (Havlíček 1982). In the uppermost part of the Kosov Formation, the quartz sandstones with shaly intercalations are replaced by siltstones and mudstones. Pale grey, often bioturbated calcareous mudstones and claystones containing a rich *Hirnantia sagittifera* Community occur near the top of the formation. The *Hirnantia* fauna, interpreted by Havlíček (1982) as representing a subtidal environment, has been found only in the eastern part of the Prague Basin. A gradual deepening of the sea seems likely in the uppermost Kosov (Hirnantian) of the Prague Basin.

The cosmopolitan *Hirnantia* fauna found in the uppermost part of the Kosov Formation permits a broad international correlation. In the Prague Basin it was first recorded at Běchovice near Prague (Marek 1963; Marek & Havlíček 1967). Later, it was found at Nová Ves, Pankrác, Řepy and Řeporyje (all within the Prague area) and near Tachlovice. All the fossiliferous localities yielded faunal associations of similar taxonomic composition, but without the depth-controlled variations of the associations reported by Brenchley & Cocks (1982) and Brenchley & Cullen (1984) from the Oslo region, Norway. Lists of the *Hirnantia* faunas from Bohemia were published by Havlíček (1982) and Štorch (1986). The graptolite *Glyptograptus bohemicus* (Marek) accompanies the *Hirnantia sagittifera* Community in Bohemia and supports the international biostratigraphic correlation of the sequence. The layer containing the *Hirnantia* fauna is separated from the first graptolitic shales by at least 0·3 m thickness of mudstone, often heavily bioturbated, with frequent limonite impregnations originating from pyrite weathering (Štorch 1986).

Ordovician–Silurian boundary and lowermost Silurian sequence

In general, sedimentation is continuous through the Ordovician–Silurian boundary in the Prague Basin, in spite of some differences between the separate sections. By using distinctive features of the boundary sequence and also the basal Silurian lithologies, the Prague Basin may be formally subdivided into five areas (Štorch 1986).

The quietest sedimentation, in probably the deepest parts of the basin, is limited to the sections along the whole south limb of the basin (South limb area—Želkovice, Vseradice, Běleč, Vočkov, Zadní Třebaň, Hlásná Třebaň, Karlík, Černošice and Velká Chuchle).

A complete succession starting with the *Akidograptus ascensus* Subzone has been preserved in all the localities (exemplified by the Karlík section, Fig. 1). Clayey shales with climacograptids and rare glyptograptids were recorded even below the first occurrences of *A. ascensus* at Želkovice and Vočkov and could represent the upper part of the *Glyptograptus persculptus* Zone. The *ascensus* Subzone is represented by clayey shales with subsidiary variable siltstones. Sandy-micaceous laminites start within the *Parakidograptus acuminatus* Subzone. The laminites disappear in the western part of the south limb at Želkovice and Všeradice in the *Cystograptus vesiculosus* Zone, and towards the east in the *Coronograptus cyphus* Zone, and sometimes they even reach up to the *Demirastrites triangulatus* Zone (Štorch 1986). In the same way, the onset horizon of siliceous shales migrates in the south limb from the *vesiculosus* Zone at Želkovice to the *Demirastrites pribyli* Zone at Černošice and Velká Chuchle. The Řepy and Běchovice sections differ in having more rapid sedimentation, giving the greatest thicknesses of graptolite zones (Řepy section, Fig. 1) in this part of the Prague Basin. The layer referred to the *per-*

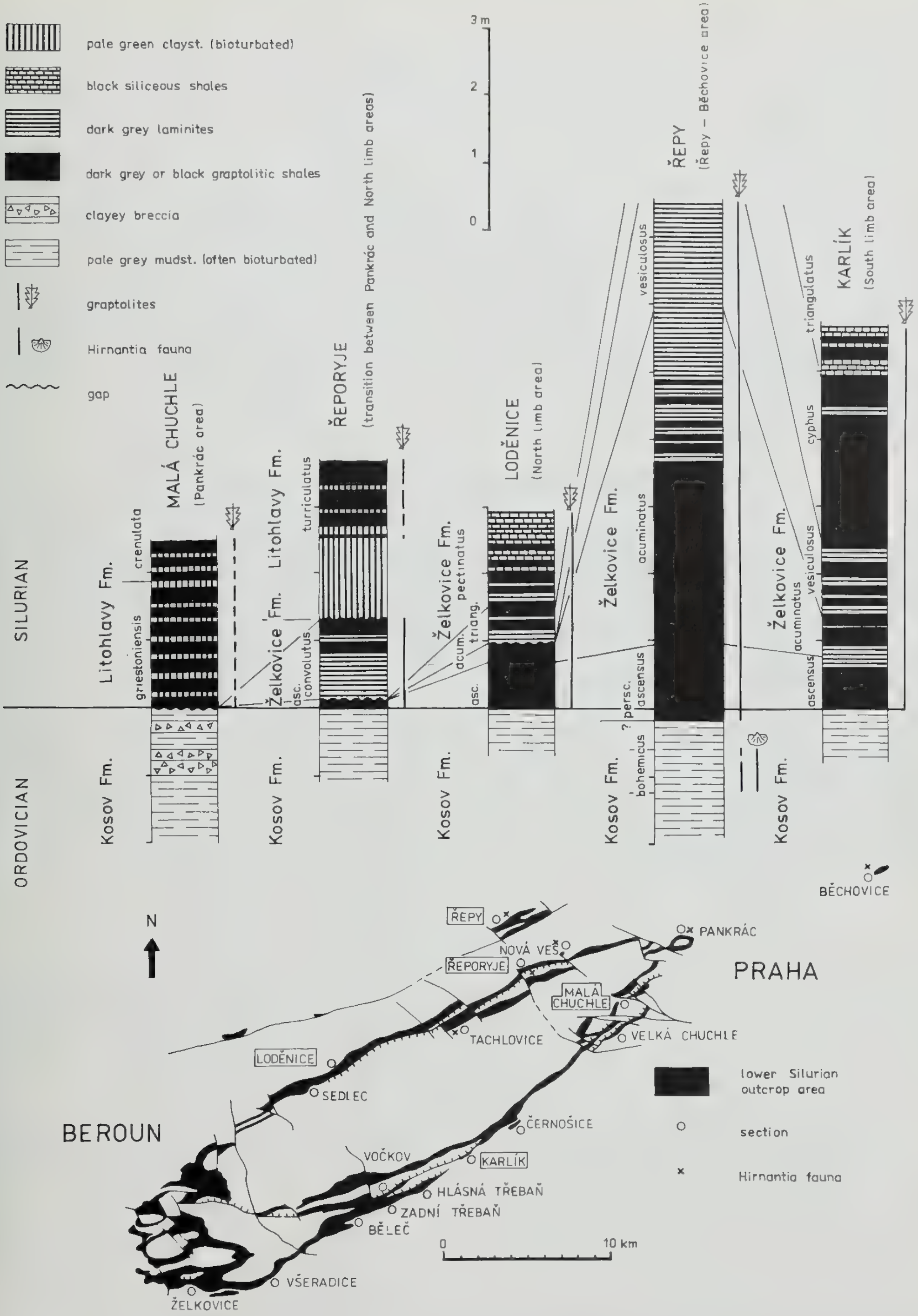
pale green clayst. (bioturbated)
black siliceous shales
dark grey laminites
dark grey or black graptolitic shales
clayey breccia
pale grey mudst. (often bioturbated)
graptolites
Hirnantia fauna
gap
3 m
2
1
0
SILURIAN
ORDOVICIAN
MALÁ CHUCHLE
(Pankrác area)
ŘEPORYJE
(transition between Pankrác and North limb areas)
LODĚNICE
(North limb area)
ŘEPY
(Řepy – Běchovice area)
KARLÍK
(South limb area)
Litohlavy Fm.
Želkovice Fm.
Kosov Fm.
crenulata
griestoniensis
turriculatus
convolutus
asc.
pectinatus
triang.
acum.
vesiculosus
acuminatus
ascensus
persc.
?
bohemicus
triangulatus
cyphus
BĚCHOVICE
N
ŘEPY
NOVÁ VES
PANKRÁC
ŘEPORYJE
PRAHA
MALÁ CHUCHLE
TACHLOVICE
VELKÁ CHUCHLE
LODĚNICE
SEDLEC
lower Silurian outcrop area
BEROUN
ČERNOŠICE
section
VOČKOV
KARLÍK
Hirnantia fauna
HLÁSNÁ TŘEBAŇ
ZADNÍ TŘEBAŇ
BĚLEČ
0
10 km
VŠERADICE
ŽELKOVICE

Fig. 1 Lithology, stratigraphy and faunal distribution in selected Ordovician–Silurian boundary sections of the Prague Basin; location of the sections.

sculptus Zone is also developed there. Laminites appear in the *acuminatus* Zone and pass up into the *vesiculosus* Zone.

Detailed studies of both biostratigraphy and lithostratigraphy (Štorch 1986) revealed that the laminites represent more condensed sedimentation than the clayey and silty shales. The onset of laminite deposition in the southern limb of the basin appears to have been synchronous with the start of the break in sedimentation in the *acuminatus* Zone in the north limb of the basin at Sedlec and Loděnice.

The longest break in sedimentation is known from the Malá Chuchle, Pankrác, Nová Ves and Tachlovice sections (Pankrác area). The topmost Ordovician mudstones are followed there by graptolitic shales of the Litohlavy Formation, with upper Llandovery graptolites of the *Monoclimacis griestoniensis* Zone. In this case, reworking possibly took place of previously deposited, incoherent, clayey and muddy sediments of the basal Silurian (*ascensus* Subzone, the lower part of the *acuminatus* Subzone), and perhaps also of the topmost Ordovician (several tens of centimetres in thickness). Near Stodůlky and Řeporyje (Řeporyje section, Fig. 1), this break in sedimentation splits into two shorter gaps. The earlier of them starts above the *ascensus* Subzone and thus supports the explanation of the break presented in different parts of the Prague Basin.

Sedimentation and assumed bathymetric changes

The Kosov Formation, which is about 100 m thick, shows sedimentation which was presumably controlled by glacio-eustatic regression. The subsequent transgression started in the uppermost Kosov and strongly accelerated at the base of the Silurian (Brenchley & Newall 1984). Considerable transgression is also documented by a decrease of the rate of sedimentation at the Ordovician–Silurian boundary. In the Prague Basin, the rate of sedimentation in the lowermost Silurian was approximately calculated (Štorch 1986) to range between 1 m and 7·5 m per 10^6 years in contrast to nearly 100 m per 10^6 years during the Kosov Series (Hirnantian). During the *acuminatus* Zone the transgression caused a further deepening of the Prague Basin and was probably the origin of a fairly intensive bottom current in the deeper central part of the linear depression of the Prague Basin. This current is considered to have caused local breaks in sedimentation, in places perhaps accompanied by mild subaquatic erosion (Štorch 1986). In the sites where this current had less erosive power, condensed sedimentation of laminites occurred, and in the quietest parts of the basin floor there were deposited siliceous shales and silty silicites ('phtanites') which first appeared in the *vesiculosus* Zone.

Stratigraphy

The *Hirnantia* fauna occurs in the upper part of the Kosov Series well above the disappearance of the *Mucronaspis* Community in the basal part of the Series. The *Hirnantia* fauna, which is accompanied by *Glyptograptus bohemicus*, can be referred to the upper Hirnantian, namely to the upper part of the *Climacograptus extraordinarius* Zone or the lower part of the *persculptus* Zone.

In Bohemia, the base of the Silurian System coincides with the base of the *ascensus* Subzone, which is defined by the first appearance of *Akidograptus ascensus* Davies (usually accompanied by *Diplograptus modestus* Lapworth). When compared with the British Isles, the base of the subzone in Bohemia is comparable to the base of the *acuminatus* Zone at the type section Dob's Linn (Williams 1983). In the Prague Basin, the base of the *ascensus* Subzone mostly corresponds to a sudden change in both the colour and the composition of the sediments, in which the pale grey bioturbated mudstones are replaced by dark grey clayey graptolitic shales. However, a low-diversity climacograptid–glyptograptid assemblage has been recorded from several localities at the base of the graptolitic shales just below the *ascensus* Subzone, which is separated by an unfossiliferous bioturbated mudstone from the *bohemicus* Zone beneath. The first assemblage of graptolitic shales below the *ascensus* Subzone is referred to the upper part of

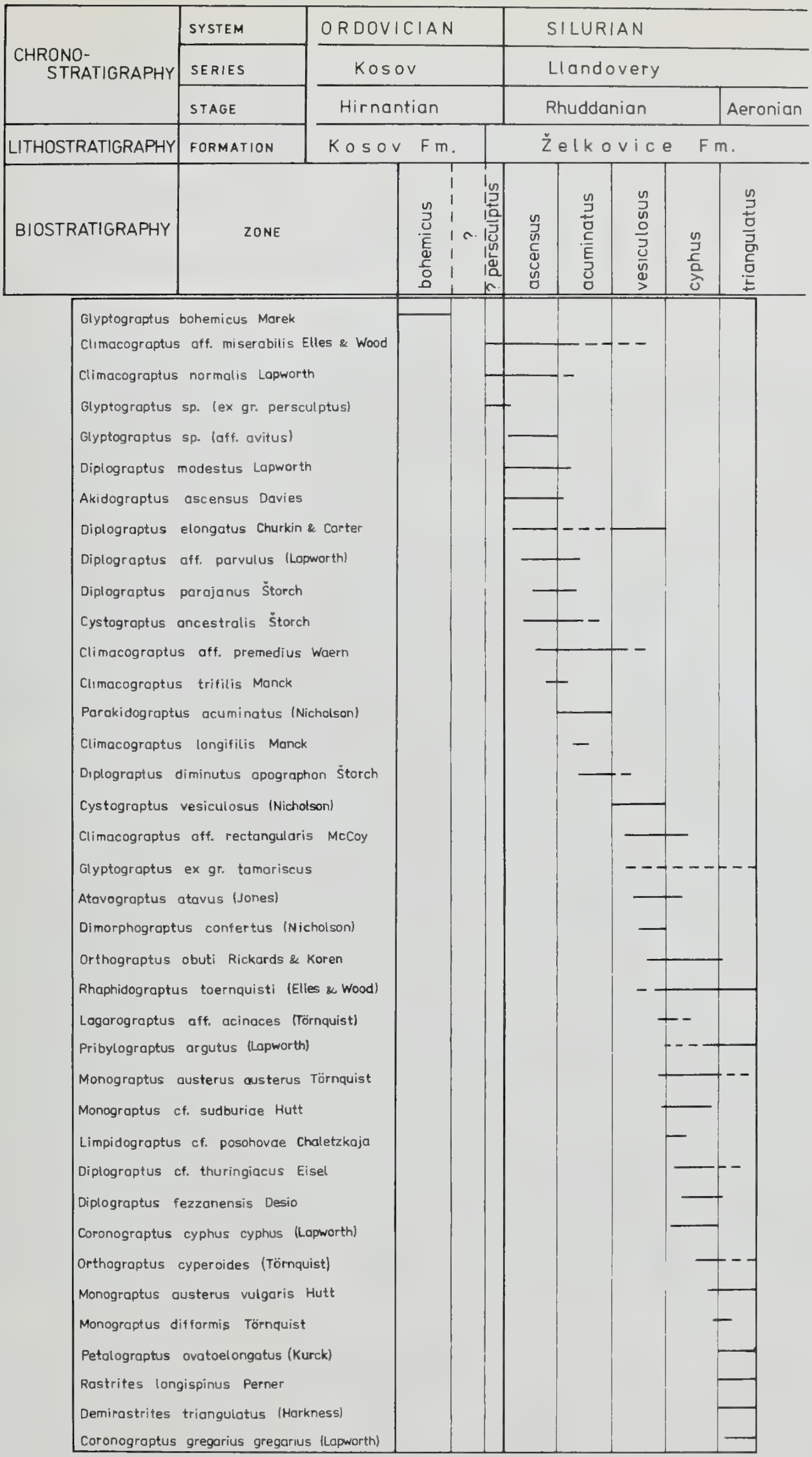

Fig. 2 Chronostratigraphy, lithostratigraphy, biostratigraphy and graptolite species ranges through the Ordovician–Silurian boundary interval in the Prague Basin.

the *persculptus* Zone, in spite of the fact that true *Glyptograptus persculptus* has not yet been found there.

The ranges of graptolites up to the base of the *triangulatus* Zone are shown in Fig. 2. The rich graptolite assemblages of the Prague Basin were briefly described by Bouček (1953), and more recently they have been described by Štorch (1986).

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The Ordovician–Silurian boundary in the Saxothuringian Zone of the Variscan Orogen

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Synopsis

In the Saxothuringian Zone of the Variscan Orogen in Thuringia, Saxonia and north Bavaria the poorly fossiliferous, thick arenaceous-argillaceous Ordovician rocks are abruptly but conformably succeeded by the very condensed sequence of Silurian–Early Devonian graptolitic alum shales and lydites beginning in both major facies with the Zone of *Akidograptus ascensus*. Below it, shaly interbeds in the uppermost Ordovician Döbra Sandstone yielded chiefly non-zonal graptolites, and in one section *Diplograptus bohemicus* about 1 m below the lithological boundary.

Introduction

The Saxothuringian and Lugian (= West Sudetic) Zones form the middle of the three major depositional and tectonic belts of the Variscan Orogen in central Europe. They constitute the metamorphic zones that are situated between the internal Moldanubian Zone (internids) and the external Rhenohercynian Zone (externids). The latter is exemplified by the Rheinisches Schiefergebirge and the Harz Mountains, in both of which the nature of the Ordovician–Silurian junction is unknown. In this paper only the type area of the Saxothuringian Zone is considered; it lies west of the River Elbe in Saxonia, Thuringia, north Bavaria and north Bohemia. Together with the Lugian Zone (situated east of the Elbe), it forms the northern part of the Bohemian Massif and is the largest outcropping fragment of the broken Variscan orogen in central Europe. In a wider palaeogeographical and geotectonical context, the Saxothuringian–Lugian Zones are part of the Mediterranean province, and of the Palaeotethys geosyncline and sea, that is the Tethys of the early and middle Palaeozoic.

In the whole of the Palaeotethys area, the Ordovician–Silurian transition is marked by a drastic change in the depositional regime. In the Saxothuringian Zone the typically 2000 m thick Ordovician, consisting of poorly fossiliferous, arenaceous-argillaceous rocks with some sedimentary iron ore bodies, is rapidly replaced by 50 m thick Silurian, which is made up almost entirely of interbedded euxinic lydites and alum shales rich in graptolites. From the middle Ludlow to the Pridoli, the graptolitic shales are interrupted by a peculiar limestone (Ockerkalk) or grey-green clay shales, both of which are poorly aerated deposits. Sedimentation of the alum shales, and regionally also of the lydites, recurred in the uppermost Silurian, and lasted well into the Lower Devonian (Lochkov).

The Silurian (and Devonian) graptolitic shales of the Thuringian type, that is alum shales and black lydites, contain large quantities of pyrite, phosphorite (in nodules and layers) and carbon (in beds, laminae and lenses). These rocks cover vast areas in the deeper parts of the Palaeotethys sea between Thuringia and north Africa. They are the result of one of the largest oceanic anoxic events in the history of the earth, both areally and temporally.

Thuringian and Bavarian Facies

In the geosynclinal Palaeozoic of the Saxothuringian Zone two major facies (or rather series of facies—Faziesreihen) are distinguished, at least in the rock-sequences from the Ordovician to the Lower Carboniferous. These are known as the 'Thuringian' and 'Bavarian Facies', but it is beyond the scope of this paper to outline their features in detail. The following points may however be made.

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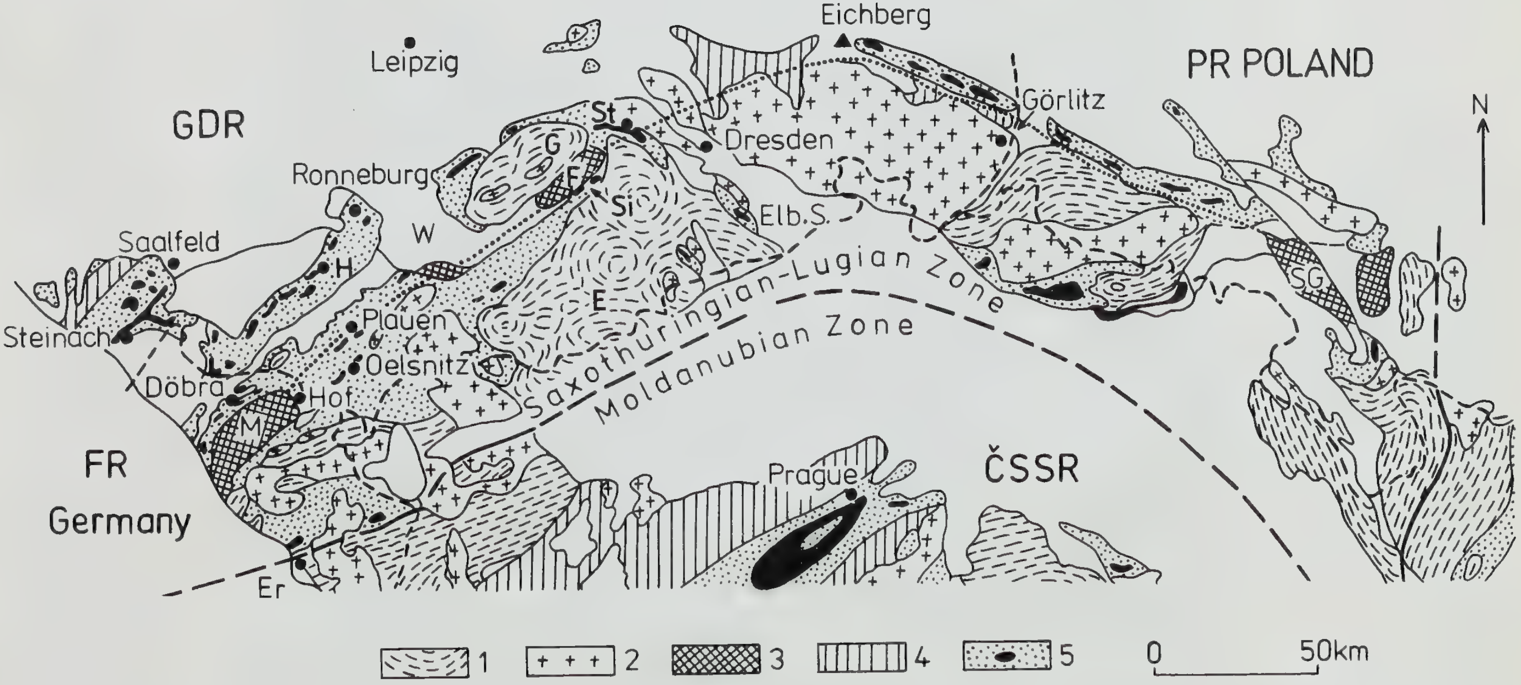
Leipzig
GDR
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Eichberg
Görlitz
Dresden
St
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F
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Ronneburg
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Saalfeld
H
Steinach
Plauen
Oelsnitz
Döbra
Hof
M
E
SG
Saxothuringian-Lugian Zone
Moldanubian Zone
FR
Germany
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N
1
2
3
4
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0
50km

Fig. 1 Geological sketch of the northern half of the Bohemian Massif. Legend: 1. Strongly metamorphosed rocks (gneisses, schists, phyllites) of Precambrian to Variscan age. 2. Major plutons (granites, granodiorites and syenites) of Precambrian to Variscan age. 3. Crystalline rocks (typically katazonal) of the Variscan 'Zwischengebirge' (Betwixt Mountains) of Münchberg (M), Wildenfels (W), Frankenberg (F) and Sowie Gory (SG = Eulengebirge). 4. Proterozoic, unmetamorphosed to weakly metamorphosed. 5. Cambrian to Devonian (black = Silurian). E = Erzgebirge. Elb.S = Elbtalschiefergebirge. Er = Erbendorf. G = Granulitgebirge in Saxonia. H = Hohenleuben. Si = Silurberg in Obermühlbach near Frankenberg. St = Starbach.

The Thuringian Facies represents a monotonous basin facies that exhibits only moderate lateral changes, if any. By contrast, the Bavarian Facies is complex. In the simplest model (Jaeger 1977: text-fig. 3) its site is depicted as a swell flanked on either side by deep furrows (deeper than the Thuringian basin). The central swell of the Bavarian Facies region is characterized by intermittent carbonate sedimentation that lasted demonstrably from the Silurian into the Carboniferous. On the swell the nature of the Ordovician–Silurian boundary is unknown. In most or all of the Saxothuringian Zone the swell-limestones are known from allochthonous blocks (olistholites) or even only from boulders, for example, the Middle to Upper Devonian stromatoporoid-coral reef-limestones at Frankenberg. The flanking depressions received non-carbonate sediments throughout their history. Typical of this Bavarian basin facies is the continuous sequence of cherts, siliceous shales and clay shales (Kieselschiefer-Fazies) spanning the long interval from the base of the Silurian to the top of the Devonian. In the Silurian interbedded graptolitic black lydites and alum shales are the typical rocks, as in the Thuringian Facies, whereas throughout most of the Devonian conodont-bearing brighter grey-green and even red cherts, siliceous shales and clay shales occur.

The region of the Bavarian Facies was, at least in its Bavarian type area, the site of large-scale basic vulcanism which lasted intermittently from the earliest Ordovician to the Carboniferous, whereas in the Thuringian Facies the geosynclinal basic vulcanism was virtually confined to a brief phase of violent eruptions and intrusions at the beginning of the Upper Devonian.

Rocks of the Thuringian Facies cover large areas in the Saxothuringian Zone. Minor occurrences are known from the southern margin of the Lugian Zone in Czechoslovakia. The Bavarian Facies rocks form a discontinuous belt that runs along the strike near the middle of the Saxothuringian Zone. They are confined to narrow strips (at the most several kilometres broad) on either side of the so-called Zwischengebirge (Betwixt Mountains) of Münchberg, Wildenfels and Frankenberg. East of the Elbe, the Bavarian Facies reappears at the Eichberg near Weissig immediately north of the plutons that build up the area between Dresden and Görlitz. From the Eichberg the Bavarian Facies can be traced through all of the Lugian Zone as far as the southern end of the Sowie Gory (Eulen-Gneis), where it is particularly well developed. Outside its main belt, the Bavarian Facies is typified by the Palaeozoic of the Elbtalschiefergebirge southeast of Dresden. The palaeogeography of the area of the Bavarian Facies may be envisaged as an island arc (the use of which term does not necessarily denote the implications of the theory of plate tectonics).

Ordovician–Silurian Boundary

At the Ordovician–Silurian boundary the distinctness of the two contrasting regional facies is particularly pronounced. In the Thuringian Facies the uppermost Ordovician is represented by the peculiar Lederschiefer, a monotonous, almost black, buff-weathering, non-bedded silty shale with high content of mica. Predominantly arenaceous rock-detritus and isolated sandstone boulders up to 30 cm across (some attaining even several metres) occur in varying quantities throughout the 250 m thick formation, for which it is noted. Whether the boulders represent glacial drop-stones or whether they originated from slumping are much debated questions. While the matrix of the Lederschiefer is barren, many boulders contain brachiopods, bryozoans, various trilobites and echinoderms, particularly loose cystoids. Most of these exotic fossils await modern expert study. Strata that compare closely lithologically with the Lederschiefer are of wide distribution in the Mediterranean province, for example in the Orea Shale in Spain.

In the uppermost two to three metres of many Thuringian sections it can be seen that the sand grains and mica flakes disappear, while many pyrite nodules appear in the shales, heralding the change to the otherwise abrupt transition to the Silurian euxinic graptolitic rocks. By contrast, the occurrence of sandstone beds in the uppermost Lederschiefer has been reported (Troeger 1959, 1960; Freyer 1959) from eastern sections (near Oelsnitz) that lie near the Bavarian Facies belt.

In view of the intense folding, sections that exhibit a tectonically undisturbed transition from the Ordovician to the Silurian are hardly to be expected between rocks with such different mechanical properties as the Lederschiefer (below) and the lydites/alum shales (above). Nevertheless, a century ago *Akidograptus acuminatus* was recovered from the basal graptolite shales at Ronneburg and Oelsnitz by Eisel. Recently Alder (1963) and Schauer (1971) found *A. ascensus* in the basal ½ m of interbedded alum shales and lydites below the *acuminatus* fauna at the Weinberg near Hohenleuben in what would appear to be the most intact boundary sections. The zone fossil is associated with *Diplograptus modestus* and several forms of *Climacograptus* (*C. medius*, *C. rectangularis*, *C. scalaris normalis* and *C. miserabilis*); there also occur unnamed climacograptids that have branched virgellae or virgellae with a distal vesicular appendage (Schauer 1971).

In the succeeding half metre, *Akidograptus acuminatus* occurs together with all the species that are already present in the *ascensus* Zone, but in addition, the highly characteristic *Climacograptus trifilis* Manck and *C. longifilis* Manck make their first appearance.

In the Bavarian basin facies the uppermost Ordovician is represented by the Döbra Sandstone. This is an almost black, fine-grained, often quartzitic sandstone with subordinate shaly interbeds, with a maximum thickness in excess of 40 m. Some sandstone beds exhibit magnificently-developed sole markings (load casts), others roll- and ball-structures. Greiling (1966: 12) interprets the Döbra Sandstone essentially as a turbidite. This peculiar rock is a characteristic formation of the Bavarian Facies, and is of wide distribution. It can be traced intermittently throughout the Saxothuringian and Lugian Zones for a total length of 400 km and it has a far greater linear extent in central Europe than the coeval Lederschiefer.

Lithologically virtually identical (Carnic Alps) or dissimilar (Kosov Quartzite in the Barrandian) sandstones occur in the same or analogous stratigraphical position in many areas of the Mediterranean province. In some regions they may range considerably higher, through much of the Llandovery, and not start until the base of the Silurian.

The Döbra Sandstone is practically unfossiliferous, except for the uppermost two metres which yielded graptolites in shaly interbeds. Stein (1965: 119; text-figs 5, 20 and others) described *Climacograptus medius*, *C. scalaris normalis*, *Diplograptus modestus*, and a single rhabdosome of *D.* cf. *persculptus* (Salter) from 1·90 m below its top at Döbra.

At the Silurberg locality in Obermühlbach near Frankenberg *Diplograptus bohemicus* (Marek) was described by Jaeger (1977) from the uppermost Döbra Sandstone. This species occurs there abundantly, but to the exclusion of other graptolites, in a layer just a few mm thick in the middle of a 0·70–0·75 m thick bed of homogeneous grey-black clay shale that underlies a prominent 30 cm thick quartzite. The latter is overlain by ½ m of platy sandstone and shale showing slickensiding, which is succeeded by 1 m of broken and mylonitized alum shales and lydites indicating a major fault that throws Ludlovian (*colonus* and *chimaera* Zone) graptolite shales against the Ordovician Döbra Sandstone. The same sequence, particularly the 0·70–0·75 m thick bed of shale and the overlying compact 30 cm sandstone bed, have been traced to the northeast as far as Starbach. This sequence is therefore shown as the typical one in Fig. 2 (right column). In the apparently undisturbed boundary section at Starbach the 30 cm thick compact sandstone bed is immediately overlain by 40 cm of weathered clay shales and siliceous shales, which in turn are succeeded by typical alum shales and lydites. Graptolites were not found in the Döbra Sandstone at other localities, nor was the occurrence of the basal Silurian graptolite zones established in this northeastern part of the Saxothuringian Zone.

The basal Silurian graptolite zones were recovered in the lowermost alum shales and lydites of the type area of the Bavarian Facies along the northwest side of the Münchberg gneis at Döbra, Förtschenbach, Ober-Brumberg and Rauhenberg (Greiling 1957, 1966; Stein 1965). Though these workers did not formally distinguish between the Zones of *A. ascensus* and *acuminatus* it would appear evident from Stein's precise documentation that the two can be differentiated. The thicknesses are approximately the same as in the Thuringian Facies, or slightly less. The associations are also the same, though the number of listed forms is somewhat smaller. *Climacograptus trifilis* and *C. longifilis* occur as frequently as in the Thuringian Facies.

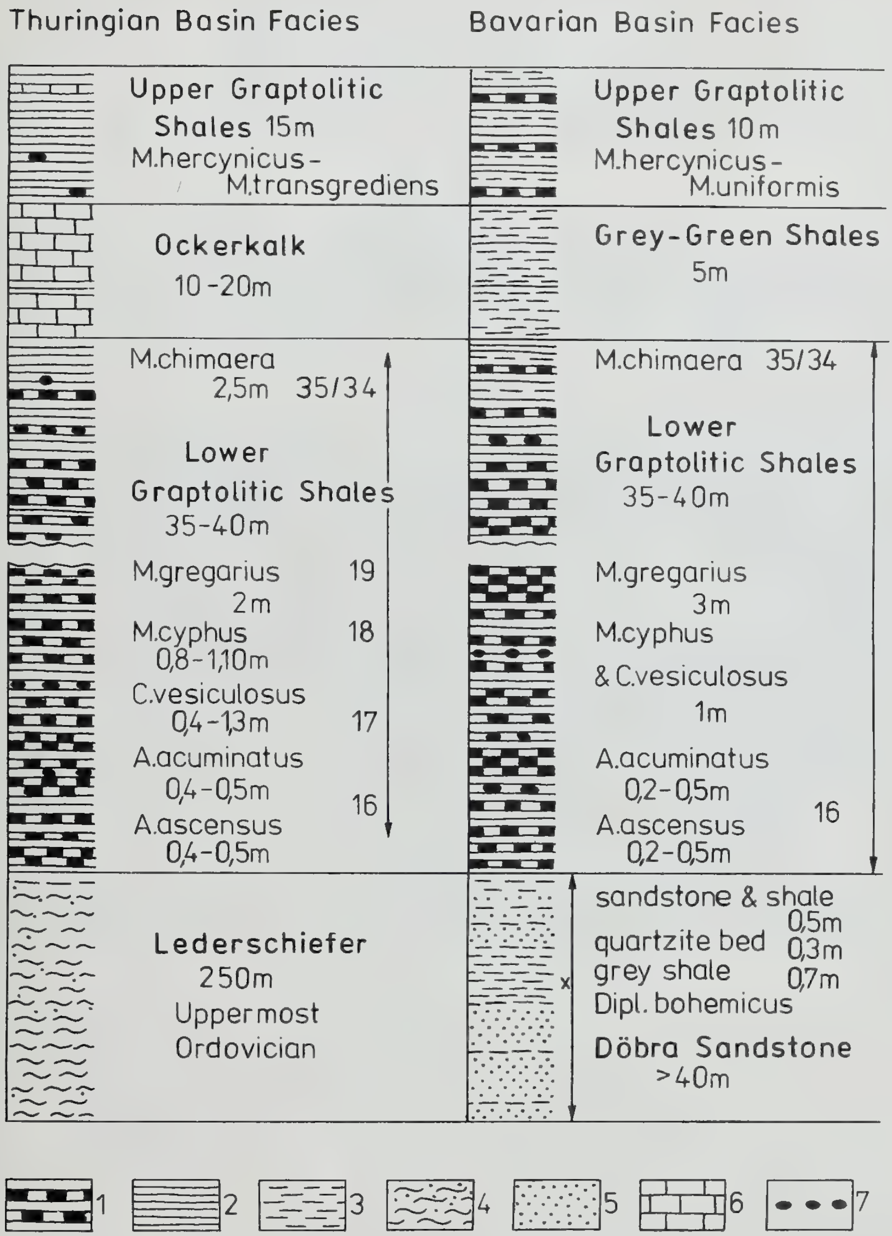
Thuringian Basin Facies
Bavarian Basin Facies
Upper Graptolitic Shales 15m
M.hercynicus - M.transgrediens
Ockerkalk
10 -20m
M.chimaera 2,5m 35/34
Lower Graptolitic Shales 35-40m
M.gregarius 2m 19
M.cyphus 0,8-1,10m 18
C.vesiculosus 0,4-1,3m 17
A.acuminatus 0,4-0,5m
A.ascensus 0,4-0,5m 16
Lederschiefer
250m
Uppermost Ordovician
Upper Graptolitic Shales 10m
M.hercynicus - M.uniformis
Grey-Green Shales
5m
M.chimaera 35/34
Lower Graptolitic Shales 35-40m
M.gregarius 3m
M.cyphus & C.vesiculosus 1m
A.acuminatus 0,2-0,5m
A.ascensus 0,2-0,5m 16
sandstone & shale 0,5m
quartzite bed 0,3m
grey shale 0,7m
x
Dipl. bohemicus
Döbra Sandstone
>40m
1
2
3
4
5
6
7

Fig. 2 Composite sections across the Ordovician–Silurian boundary in the Thuringian (left) and Bavarian basin facies (right). Legend: 1. Lydites (black layered cherts). 2. Alum shales. 3. Grey to green argillaceous shales. 4. Homogeneous non-bedded silty shales. 5. Arenaceous rocks. 6. Limestones. 7. Phosphoritic nodules.

Two points of general interest may be made. Firstly, in the Saxothuringian Zone, the change from the Ordovician Lederschiefer and Döbra Sandstone, respectively, to the Silurian graptolitic rocks takes place at the base of the Zone of *A. ascensus* and above beds with *D. bohemicus* which have only been found in one section of the Bavarian Facies. Secondly, in the Saxothuringian region, *A. ascensus* and *A. acuminatus* indicate two successive graptolite zones, as in the Barrandian area and southern Spain (Jaeger & Robardet 1979: 693, section 4), although *A. ascensus* ranges into the *acuminatus* Zone, and in Sardinia even into the next higher Zone of *Cystograptus vesiculosus* (Jaeger 1976: pl. 3, fig. 7).

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The Ordovician–Silurian boundary in the Carnic Alps of Austria

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Synopsis

Although the Ordovician–Silurian boundary is represented in some places by a considerable unconformity in the Carnic Alps, in other sections a *Hirnantia* fauna in the Plocken Formation and possibly *persculptus* Zone graptolites are succeeded by the Bischofalm facies which in places has yielded graptolites of the *acuminatus* Zone. The shallow-water facies and unconformities at and near the boundary were partly caused by the global eustatic fall and rise in sea level and partly by Caledonian tectonic activity.

Introduction

The long geological history of the Carnic Alps of Austria and northern Italy lasts from the late Ordovician to middle Triassic times. For many years in this region several sections which cross the Ordovician–Silurian boundary and represent different environmental settings have been well known. Based on earlier studies by Gaertner (1931), Walliser (1964), Flügel (1965), Serpagli (1967), Schönlaub (1969, 1971), Vai (1971), and Jaeger *et al.* (1975), a brief summary of knowledge of this interval up to the year 1975 was submitted and published in an earlier circular of the Ordovician–Silurian Boundary Working Group.

Based on the final decision of the Commission on Stratigraphy that the base of the Silurian System shall be at the base of the *A. acuminatus* Biozone, the present paper revises the stratigraphy of the boundary beds in the Carnic Alps. In addition new field data are presented and summarized in this updated version of previous reports. I acknowledge the help of H. Jaeger, Berlin, and R. Schallreuter, Hamburg, who kindly provided unpublished data on graptolites and ostracods.

Upper Ordovician sediments and stratigraphy

All known late Ordovician and early Silurian boundary sequences show clear evidence of a regressive–transgressive relationship. Except for one section representing the deep water 'Bischofalm graptolite facies', for which, however, biostratigraphical data are missing for the late Ordovician, the lithology and faunal composition in the upper Ordovician reflect a stable environment of shallow to moderate depths with a considerable clastic influx in the Caradoc Stage. During this time the fossiliferous Uggwa Shales, up to 100 m thick, were deposited. They comprise sandy shales and pass laterally into greenish and brownish mudstones and siltstones, the latter being widely distributed in the Central Carnic Alps in the surroundings of Plöckenpass and Lake Wolayer. In contrast to the typical Uggwa Shales, in these beds only very few fossils occur. This shale and siltstone sequence grades laterally and in part also vertically into 40–60 m of thick well-bedded and locally cross-bedded sandstones also known as the Himmelberg Sandstone. Fossils, if any, are extremely rare except for the under- and overlying strata which suggest a late Caradoc age for this unit. Hence, this sandstone is in part equivalent to the Uggwa Shale, which is also supported by field observations. The fauna of the clastic upper Ordovician sequence is dominated by bryozoans and less frequently brachiopods, trilobites, gastropods and cystoids occur. According to Vai (1971) and Havlíček *et al.* (1987), this fauna suggests a close relationship to middle Caradoc sequences of Sardinia and other regions of southern Europe as well as to Bohemia.

The Caradoc Uggwa Shale and its equivalent, the Himmelberg Sandstone, are overlain by distinctive limestones of Ashgill age. Two lithological types are developed in the Carnic Alps,

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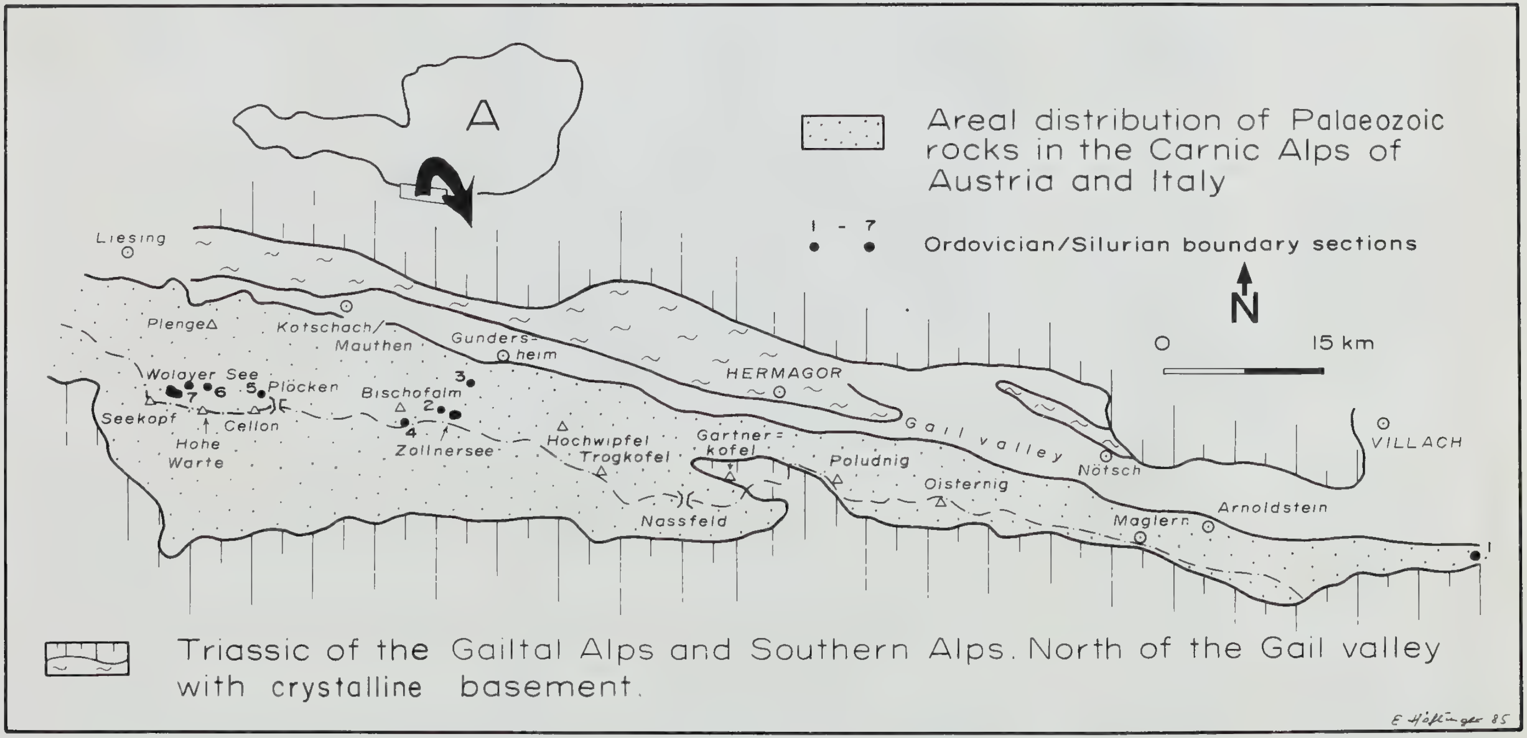

A
Areal distribution of Palaeozoic rocks in the Carnic Alps of Austria and Italy
1 - 7
Ordovician/Silurian boundary sections
N
0
15 km
Liesing
Plenge
Kotschach/Mauthen
Gundersheim
HERMAGOR
Gail valley
Nötsch
VILLACH
Wolayer See
Plöcken
Bischofalm
Seekopf
Cellon
Hohe Warte
Zollnersee
Hochwipfel
Trogkofel
Gartner=kofel
Nassfeld
Poludnig
Oisternig
Maglern
Arnoldstein
1
2
3
4
5
6
7
Triassic of the Gailtal Alps and Southern Alps. North of the Gail valley with crystalline basement.
E Höfinger 85

Fig. 1 Map of the Carnic Alps area in south Austria and north Italy.

the first being the nodular Uggwa Limestone and the other its lateral equivalent, the coarse-grained biodetrital Wolayer Limestone. The Uggwa Limestone represents a quiet water shelf environment and contains relatively abundant microfossils, for example conodonts, ostracods and foraminiferans, but also a few trilobites, bryozoans, brachiopods and cephalopods. Yet age assignments within the Ashgill are not precisely known except for its upper part, in which the *Hirnantia* fauna is found.

The second type, the Wolayer Limestone, comprises biodetrital cystoid-bearing light grey limestones which may be up to 18 m thick, three times as much as the Uggwa Limestone. Its palaeogeographic setting suggests carbonate mud mounds on the outer shelf surrounded by rather uniform and more widely distributed shelf carbonates of the Uggwa Limestone. There is no indication of close proximity to a land area for either type. In the Carnic Alps lateral changes between the two limestone types can occur over a few km in the same tectonic unit. In other places they are tectonically separated. As shown in the diagrams (Figs 1, 2) the individual boundary sections exhibit significant differences in thickness and lithology, as far as the latest Ordovician is concerned.

The Boundary Beds

At the top of the Ordovician sequence in the Carnic Alps a widespread sandy facies occurs, the so-called Plöcken Formation. In the old literature this horizon was termed ‘Untere Schichten’. It succeeds the Uggwa Limestone but is missing at the top of the coeval Wolayer Limestone (see below). Reinvestigation of the Plöcken Formation indicates that it represents a regressive sequence starting with offshore shaly mud intercalations in the uppermost Uggwa Limestone and above, and developing into shoreface calcareous sands. In these beds contorted deformation structures are very common. In the lower parts they are associated with channel fillings of coarse bioclastic material.

The Hirnantian fauna which first occurs in laminated greenish-greyish mudstones overlying the Uggwa Limestone at Cellon shows evidence of transportation. The same is true for the Hoher Trieb section east of Cellon (Figs 4E, 4F). The poorly sorted, mostly disarticulated fossil debris occurs in several layers. They are characterized by internal erosional surfaces, small-scale channelling, reworking of sediment, bioturbation with subsequent infilling of fossils, and pronounced load deformation structures. Higher up in the sequence channelling and reworking of the sediment increase, although laminated mudstones are here less abundant. Usually channels are connected with contorted beds the thickness of which is usually between 10 and 20 cm but which may reach 60 cm.

The channel filling consists of coarse-grained bioclastic limestones which cut into the underlying mudstones and shales. Fossils include representatives of the *Hirnantia* fauna (mainly brachiopods and trilobites), pyritized ostracods and spicules. According to Jaeger *et al.* (1975) and Schönlaub (1980: fig. 27 and 1985: fig. 25a) the following taxa have been found in the latest Ordovician Hirnantian Stage:

Kinnella kielanae (Temple)
Rafinesquina sp.
Clarkeia sp.
Hirnantia sagittifera (M‘Coy)
Dalmanella testudinaria (Dalman)
Cryptothyrella sp.
Dalmanitina mucronata (Brongniart)
Icriodella sp.
Quadrijugator harparum (Troedsson)
Scanipisthia rectangularis (Troedsson)
Eocytherella troedssonia Bonnema
Dornbuschia ostseensis Schallreuter

At Cellon (Fig. 3) and Hoher Trieb (Figs 4E, 4F) the channels are connected or grade into contorted beds composed of less pure limestones. They are irregularly coloured brownish and greyish marls with floating brachiopod valves and loosely packed matrix-supported subangular clasts of different rock types including carbonates of different size up to 20–30 cm in diameter, sandstone pebbles, shales or small black phosphorite nodules. At the Nölblinggraben section at the base of the Plöcken Formation there is even a layer with clasts of granitic composition (Schönlaub & Daurer 1977).

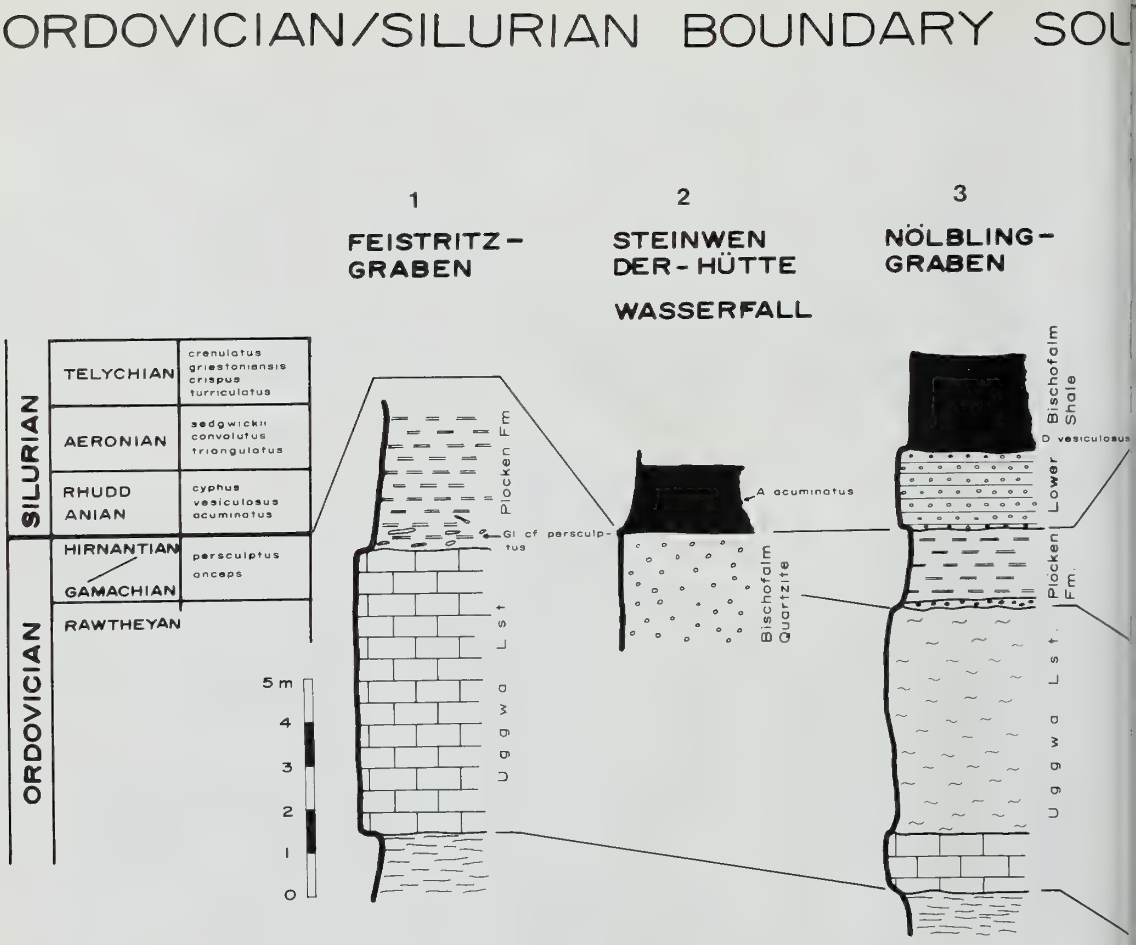
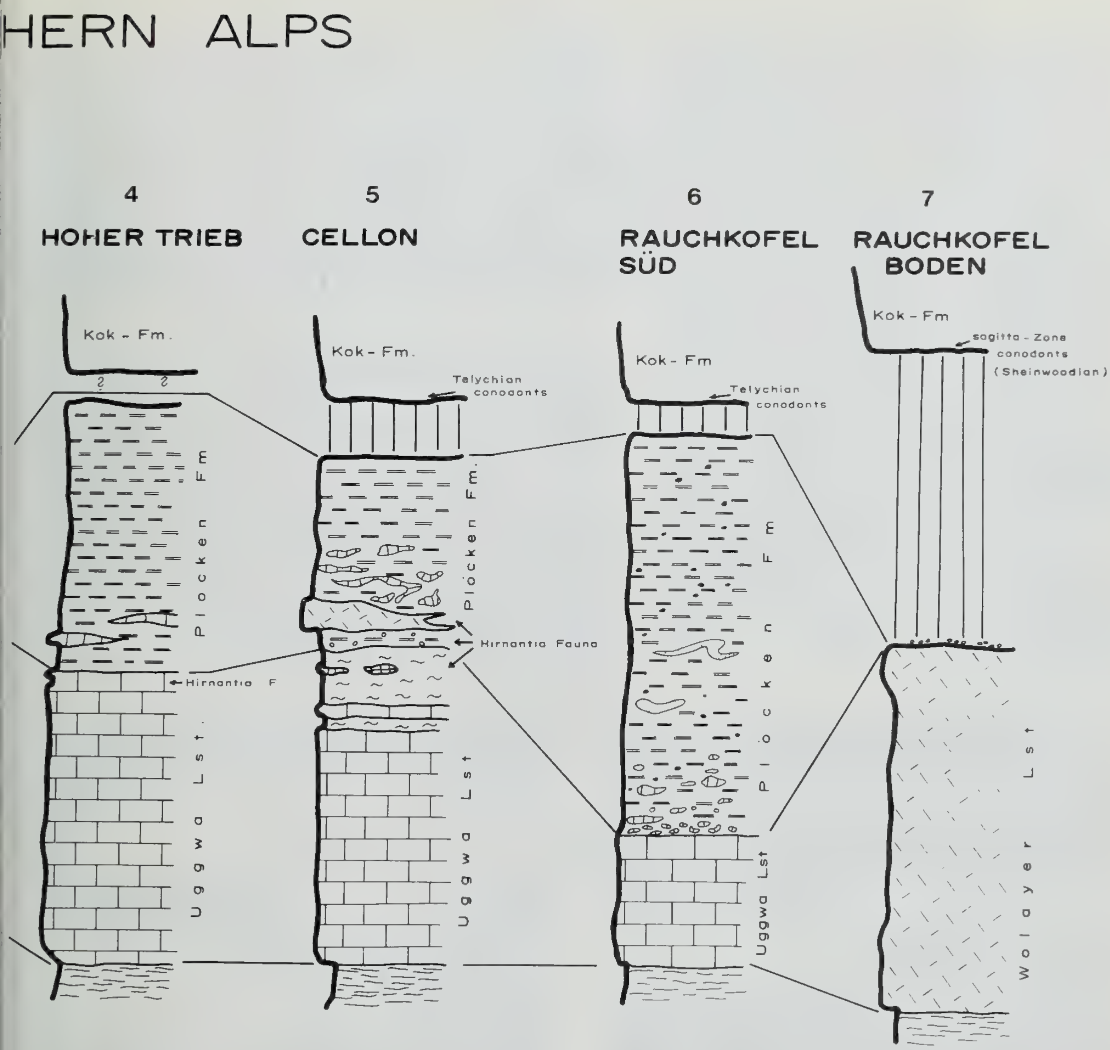

ORDOVICIAN/SILURIAN BOUNDARY SOL
1
FEISTRITZ-GRABEN
2
STEINWENDER-HÜTTE WASSERFALL
3
NÖLBLING-GRABEN
SILURIAN
ORDOVICIAN
TELYCHIAN
AERONIAN
RHUDDANIAN
HIRNANTIAN
GAMACHIAN
RAWTHEYAN
crenulatus
griestoniensis
crispus
turriculatus
sedgwickii
convolutus
triangulatus
cyphus
vesiculosus
acuminatus
persculptus
onceps
5 m
4
3
2
1
0
Plöcken Fm
Gl cf persculptus
Uggwa Lst
A acuminatus
Bischofalm Quartzite
Bischofalm Shale
D vesiculosus
Lower
Plöcken Fm.
Uggwa Lst.

Fig. 2 Comparative sections through the Carnic Alps near the Ordovician–Silurian boundary.

The Plöcken Formation has a thickness of between 1·5 and 9 m, the latter occurring on the southern slope of Mount Rauchkofel. Based on the occurrences of the *Hirnantia* faunal assemblage at the Cellon, Hoher Trieb and Uggwa sections, a late Ashgill age, i.e. the Hirnantian Stage, is deduced for the Plöcken Formation. This is in agreement with earlier reports of *Glyptograptus* cf. *persculptus* (Salter) from the 'Feistritzgraben' section in the Western Karawanken Alps (Jaeger *et al.* 1975). We correlate this level with the basal Plöcken Formation in the Carnic Alps, although the lithologies are slightly different.

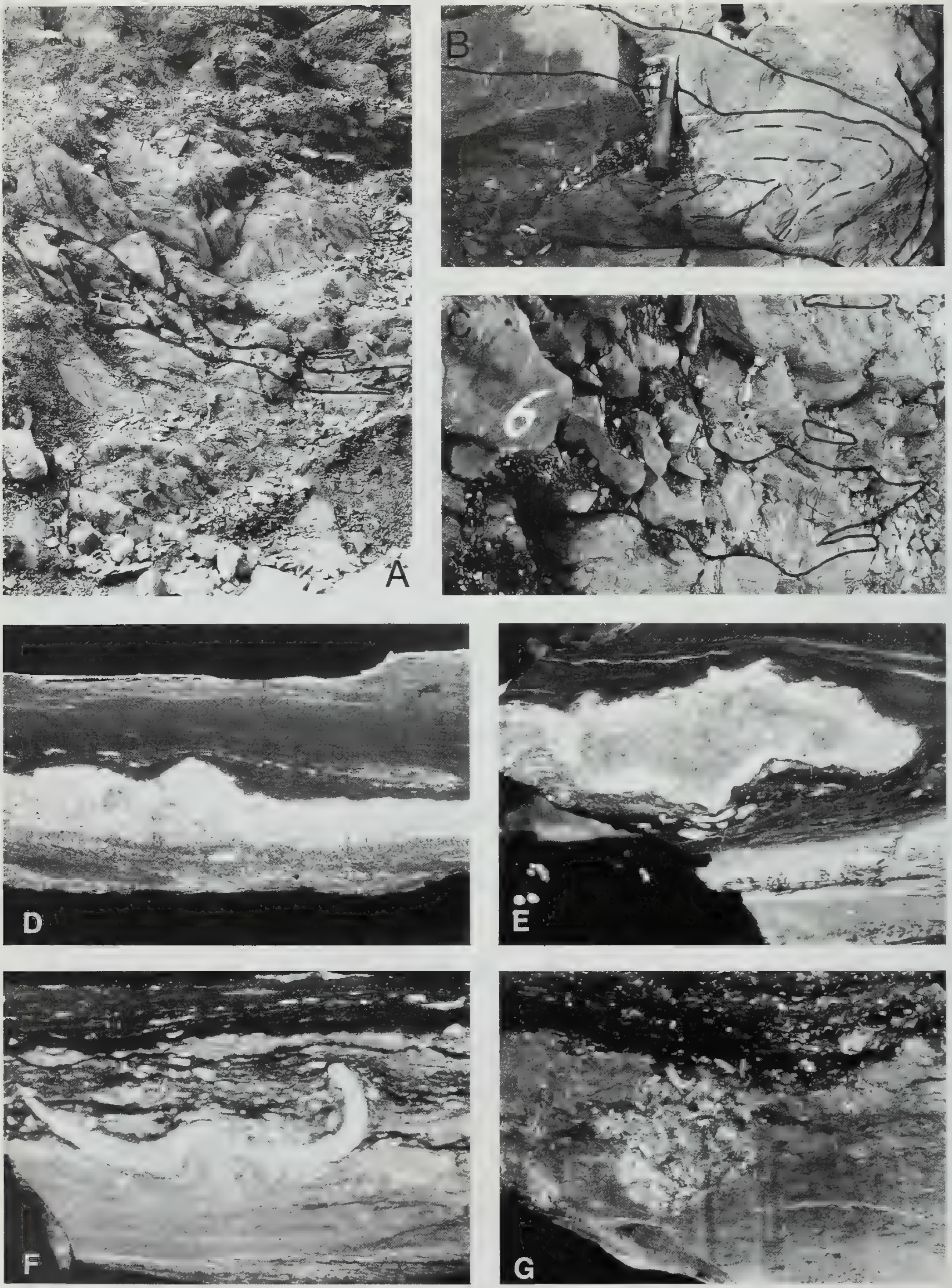
The Base of the Silurian

On the carbonate shelf which was already shallow in pre-Hirnantian times the shallow water carbonate facies was re-established in the Silurian. However, in this facies disconformities with distinct karst surfaces are widely developed and depositional hiatuses are well known. The relief may be several cm or more. In particular this phenomenon can be seen on top of the carbonate mounds of the Wolayer Limestone which apparently became subaerially exposed from the


HERN ALPS
4
HOHER TRIEB
5
CELLON
6
RAUCHKOFEL SÜD
7
RAUCHKOFEL BODEN
Kok - Fm.
?
?
Plöcken Fm
Hirnantia F
Uggwa Lst.
Kok - Fm.
Telychian conodonts
Plöcken Fm.
Hirnantia Fauna
Uggwa Lst
Kok - Fm
Telychian conodonts
Plöcken Fm
Uggwa Lst
Kok - Fm
sagitta - Zone conodonts (Sheinwoodian)
Wolayer Lst

latest Ordovician to the middle or even upper Silurian (see Fig. 2, section no. 7). In other sequences stratigraphical gaps are of shorter duration. In any case there is an abrupt upward transition from the Hirnantian Plöcken Formation to either cephalopod limestones of the Kok Formation or to the uniform dark grey graptolitic shales of the basal Silurian Bischofalm facies.

According to unpublished new data of H. Jaeger (cited by Schönlaub, 1985: 78) in the Carnic Alps the graptolite facies starts in the *A. acuminatus* Biozone. At the 'Steinwenderhütte-Wasserfall' locality the graptolitic shales succeed the greyish Bischofalm Quartzite. At other places, for example at Nölblinggraben, *D. vesiculosus*, the index graptolite of the lower Silurian graptolite zone 17, has been reported overlying an almost 2 m thick quartzitic rock. Due to the lack of fossils the stratigraphical relationship between the two quarzitic members is yet poorly understood. They may represent fan deposits of different ages, the lower one being deposited in basin areas of the Hirnantian low sea level stand and the latter at or near the beginning of the transgressive graptolite sequence at the presumed base of the Rhuddanian Stage. In either case, in this part of the Carnic Alps an almost complete succession of strata across the Ordovician–Silurian boundary can be assumed.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Conclusion

The Ordovician–Silurian boundary beds in the Carnic Alps reflect a regressive–transgressive cycle. Alongside probably continuous sedimentation across the systemic boundary in sections representing deeper environments, in the shallow carbonate shelf areas stratigraphical gaps are very common. This relation is in accordance with data from other regions in the world. However, this event was not solely caused by worldwide eustatic changes of sea level attributed to the famous glacial event in the southern hemisphere. Vertical block movements of Caledonian age also affected the Carnic Alps in the late Ordovician and, consequently, were also responsible for differences in thickness of closely-related sections as well as for greatly differing facies that developed in the Silurian after a less pronounced facies pattern in the Ordovician.

Fig. 3 Ordovician–Silurian boundary beds at the Cellon section in the central Carnic Alps of Austria. A: Cellon section, lower part showing Uggwa Limestone in the lower portion and Plöcken Formation above. Indicated is a coarse grained channel filling limestone bed at the base of the Plöcken Formation. B: Detail from A in the upper portion of the Plöcken Formation showing a multilayered fold. C: Detail from A. Coarse-grained limestone bed at the base of the Plöcken Formation (no. 6 is a reference point of O. H. Walliser's conodont-based collection). D: Internal erosional surface in the uppermost Uggwa Limestone Formation at level no. 5 of Walliser (1964). Length of the cut approx. 4 cm. E: Reworked limestone clast at the same horizon as Fig. D. Long axis approx. 3·5 cm. F: Fossil debris representing components of the *Hirnantia* fauna in the uppermost Uggwa Limestone Formation at horizon no. 5 of Walliser (1964). Width of the brachiopod valve is 3 cm. G: Same horizon as Figs D–F showing bioturbation and infilling at an internal erosional surface in mudstones. Length of the cut approx. 4 cm.

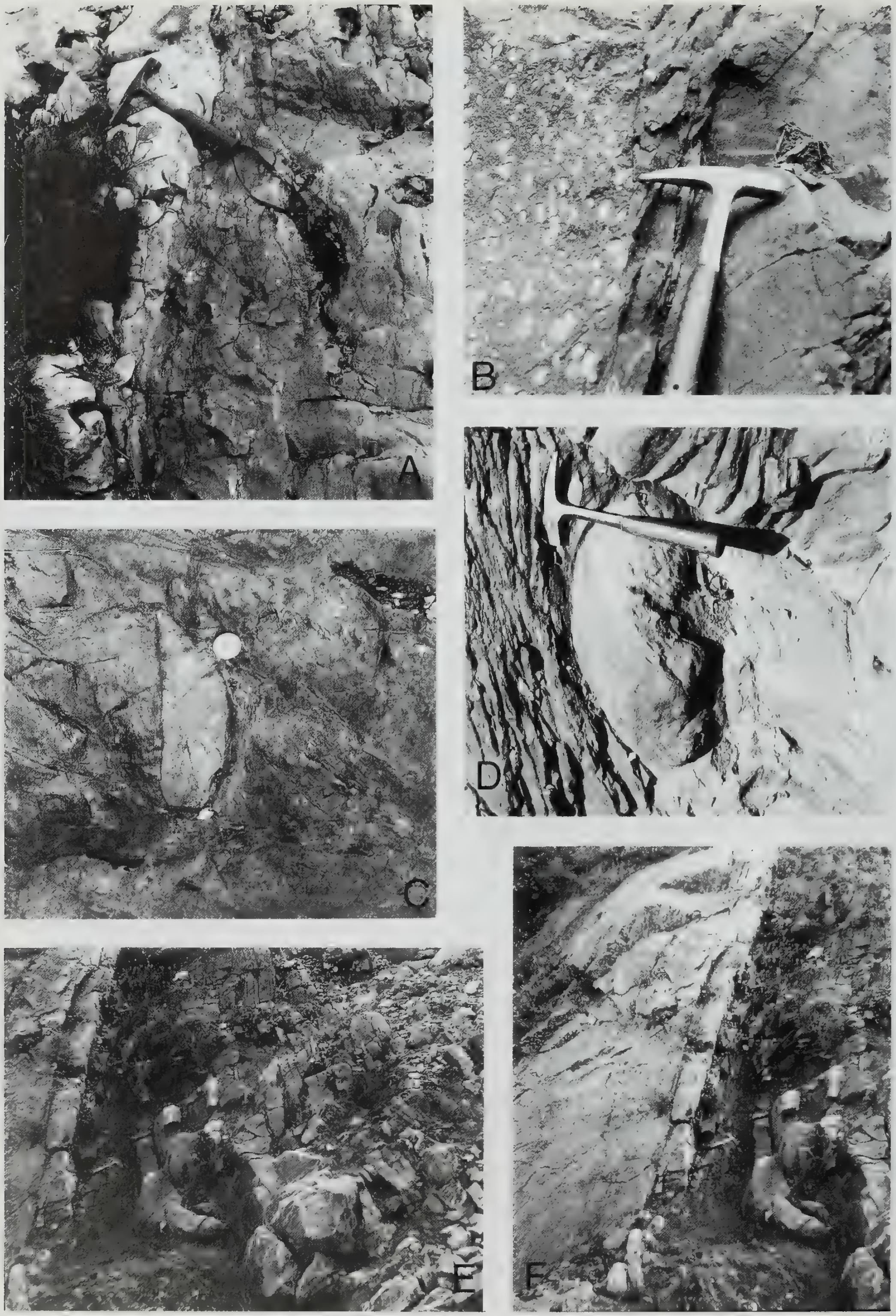
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Fig. 4 Ordovician–Silurian boundary sections at Rauchkofel-Boden, Rauchkofel-Süd and Hoher Trieb in the central Carnic Alps. A: Rauchkofel-Boden section, disconformity between the Ashgill Wolayer Limestone (left) and the darker cephalopod-bearing Kok Formation (right). At the base of the latter *sagitta*-Zone conodonts of middle Wenlock age occur. B: Rauchkofel-Süd section showing contact between the nodular Uggwa Limestone (left) and the overlying Plöcken Formation (right). C, D: Reworked limestone clasts containing an *Amorphognathus ordovicicus* conodont fauna in the lower part of the Plöcken Formation at the Rauchkofel-Süd section. E, F: Hoher Trieb section. Uggwa Limestone (left) and basal part of the Plöcken Formation (right). Note channel filling coarse-grained bioclastic bed near the base of the Plöcken Formation. This bed contains representatives of the *Hirnantia* fauna (*Hirnantia sagittifera, Dalmanella testudinaria, Kinnella kielanae, Cryptothyrella* sp. and also *Clarkeia* sp.).

The Ordovician–Silurian boundary in China

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Synopsis

After a general account of the Chinese graptolite zones about the boundary, a précis is given of the Chinese type section for the boundary, at Wangjiawan, which includes the faunal characteristics. It is followed by similar details for nine other major Chinese sections and a synthesis of the biofacial types. After a discussion of correlation problems about the boundary, it is concluded that the *ascensus* Zone of some European sections is equivalent to the Chinese *persculptus* Zone, and that the base of the Silurian is best taken above the *bohemicus* Zone and its correlatives, the *Hirnantia–Dalmanitina* fauna.

Introduction

Ordovician and Silurian strata are well developed in China. Many Ordovician–Silurian boundary sections have been defined in the Yangtze Region (or the Central China region) where the Ordovician and Silurian consist of platform deposits. These sections are small in thickness and rich in fossils, mainly graptolites, known as the Ashgill Wufeng Formation and the early Llandovery Lungmachi Formation. Between these two formations there is usually a thin bed of shelly facies, namely the *Hirnantia–Dalmanitina* bed (HD) or the Kuanyinchiao bed. The graptolite sequences of the Wufeng Formation and the Lungmachi Formation are quite complete, and thirteen graptolite zones have been established in descending order as follows:

Lungmachian:	L_7	*Monograptus sedgwickii* Zone
	L_6	*Demirastrites convolutus* Zone
	L_5	*Demirastrites triangulatus* Zone
	L_4	*Pristiograptus cyphus* Zone
	L_3	*Orthograptus vesiculosus* Zone
	L_2	*Parakidograptus acuminatus* Zone
	L_1	*Glyptograptus persculptus* Zone
Wufengian:	W_6	*Diplograptus bohemicus* Zone
	W_5	*Paraorthograptus uniformis* Zone
	W_4	*Diceratograptus mirus* Zone
	W_3	*Tangyagraptus typicus* Zone
	W_2	*Dicellograptus szechuanensis* Zone
	W_1	*Amplexograptus disjunctus yangtzeensis* Zone or *Pleurograptus lui* Zone

The establishment of the Wufengian and Lungmachian graptolite zones is of great importance in stratigraphical correlation and in the determination of the exact position of the *Hirnantia–Dalmanitina* bed (HD). The HD bed is underlain by beds of varying age from the *Tangyagraptus typicus* Zone (W_3) to the lower part of the *Diplograptus bohemicus* Zone (W_6^1) in different localities. By comparison, the earliest Silurian shelly facies, known as the '*Eospirigerina*' bed or the Wulipo bed, has a less wide distribution and its upper limit varies in different places and may reach as high as the *Pristiograptus cyphus* Zone (L_4). The relationship between the Ordovician–Silurian boundary graptolite zones and the shelly beds may be shown in Table 1.

As shown in the table, the Ordovician–Silurian boundary should be drawn between the *Diplograptus bohemicus* Zone (W_6)/*Hirnantia–Dalmanitina* bed and the *Glyptograptus persculptus* Zone (L_1)/'*Eospirigerina*' bed. The striking faunal changes from the topmost Ordovician (W_6) and the lowermost of the Silurian (L_1) support this assertion. Therefore, nearly all

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Table 1 A correlation between the graptolite and shelly sequences across the Ordovician–Silurian boundary.

L_4	*Pristiograptus cyphus*			
L_3	*Orthograptus vesiculosus*			Wulipo bed
L_2	*Parakidograptus acuminatus*			
L_1	*Glyptograptus persculptus*		'*Eospirigerina*' fauna	
W_6	*Diplograptus bohemicus*	upper (W_6^2) lower (W_6^1)	*Hirnantia–Dalmanitina* fauna (HD)	Kuanyinchiao bed
W_5	*Paraorthograptus uniformis*			
W_4	*Diceratograptus mirus*			
W_3	*Tangyagraptus typicus*			

geologists and palaeontologists in China agree that the Ordovician–Silurian boundary should be placed between the *D. bohemicus* Zone (W_6) (or the *Hirnantia–Dalmanitina* bed (HD)) and the *G. persculptus* Zone (L_1).

Description of the Ordovician–Silurian boundary sections

In 1983 the writer reviewed sixteen Ordovician–Silurian boundary sections distributed in four stratigraphical regions and described nine sections in the Yangtze Region in detail. In recent years, some sections have been revised and some new sections recognized. There are 33 well defined Ordovician–Silurian boundary sections distributed in four regions of China. Among them, 26 are in the Yangtze Region, three in the Xizang (Tibet)–W. Yunnan Region, two in the Zhujiang Region (S. China Region) and one in the Northwest Region, as shown in the map (Fig. 1). In the northernmost region, the Ordovician–Silurian strata are very thick, complicated in structure and fossils are rare, and thus no ideal Ordovician–Silurian boundary section has been found in this region. There are no Silurian deposits in the Huanghe Region (N. China Region).

In the present paper, the type section, the Wangjiawan section of Yichang, W. Hubei, and nine selected sections are described as follows.

1. The Wangjiawan Ordovician–Silurian Boundary section is the type section in China. In 1982, this section was restudied by Mu En-zhi, Zhu Zhao-ling, Lin Yao-kun, Zou Xi-ping, Wu Hong-ji, Chen Ting-en, Geng Liang-yu and Dong Xi-ping. The section is as follows (after Mu *et al.* 1984).

Lower Silurian Lungmachi Formation (basal part):

15. Black argillaceous shale weathered greyish black, yielding (ACC768) *Orthograptus vesiculosus* (Nicholson), *Climacograptus normalis* Lapworth and *C.* cf. *medius* Törnquist more than 1·0 m

14. Brownish-grey siliceous shale intercalated with black shale, with 7 siliceous beds in a distance of 20 cm, yielding (ACC767) *Parakidograptus acuminatus* (Nicholson), *Climacograptus normalis* Lapworth, *C. sinitzini* (Chaletzkaya), *Glyptograptus tamariscus magnus* Churkin & Carter and *Paraorthograptus* sp. 0·60 m

13. Black shale with (ACC766) *Parakidograptus acuminatus* (Nicholson), *Climacograptus bicaudatus* Chen & Lin, *C. normalis* Lapworth, *C. angustus* Perner and *C. sinitzini* (Chaletzkaya). 0·35 m

12. Black shale with sandy shale (0·15 m thick) in the upper part, weathered greyish black, containing (ACC765) *Akidograptus ascensus* Davies, *Glyptograptus sinuatus* (Nicholson), *G. tamariscus magnus*

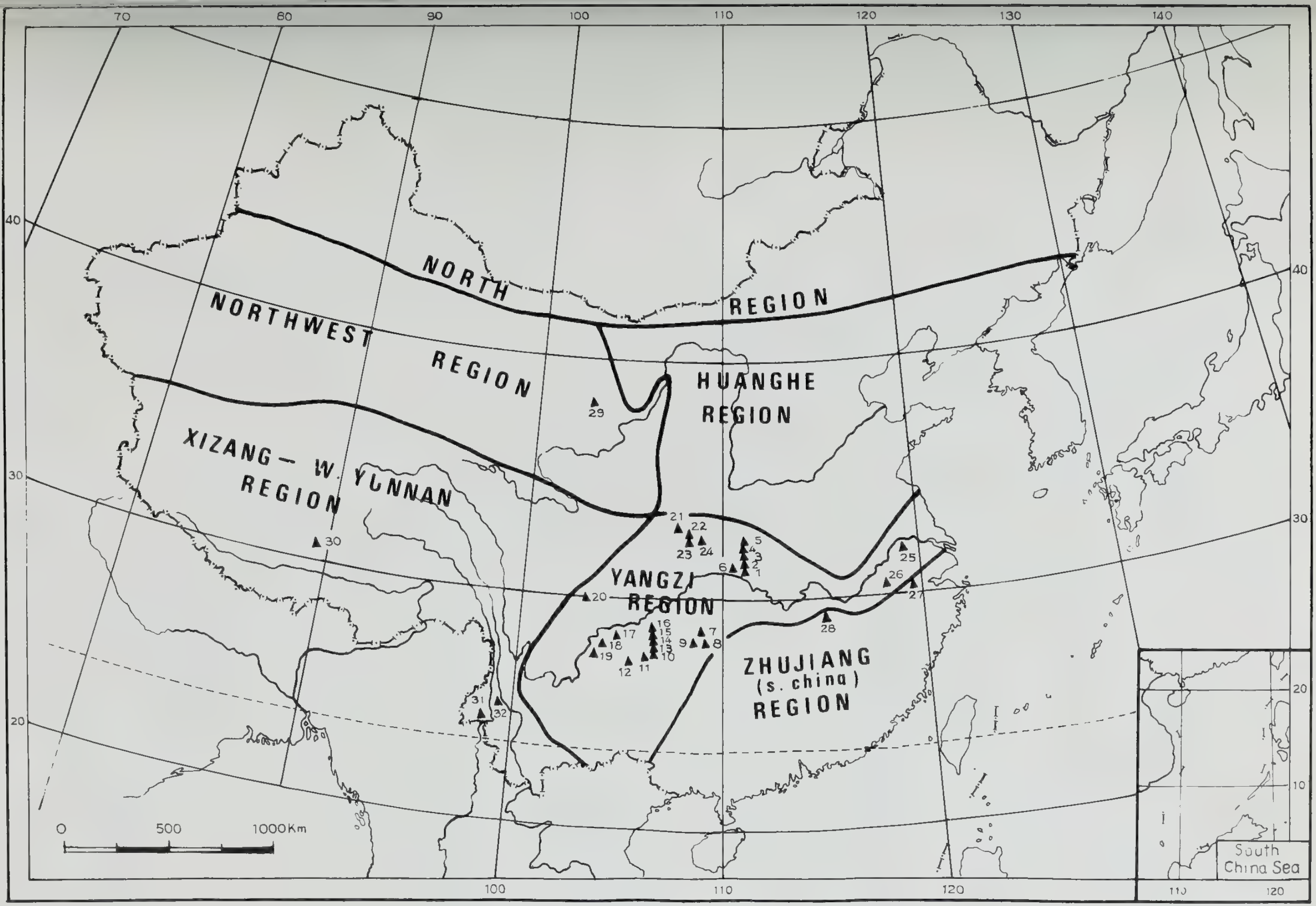
NORTH REGION
NORTHWEST REGION
HUANGHE REGION
XIZANG – W. YUNNAN REGION
YANGZI REGION
ZHUJIANG (s. china) REGION
South China Sea
0
500
1000Km
70
80
90
100
110
120
130
140
40
30
20
10

Fig. 1 Ordovician–Silurian boundary outcrops in China. 1 Huanghuachang, Yichang, W. Hubei; 2 Fenxiang, Yichang; 3 Wangjiawan, Yichang; 4 Tangya, Yichang; 5 Maliangping, Baokang, W. Hubei; 6 Xintan, Zigui, W. Hubei; 7 Xiushan, SE Sichuan; 8 Songtao, NE Guizhou; 9 Ganxi, Yanhe, NE Guizhou; 10 Donggonsi, Zunyi, N. Guizhou; 11 Renhuai, N. Guizhou; 12 Yanzikou, Bijie, NW Guizhou; 13 Huanghuayuan, Tongzi, N. Guizhou; 14 Laingfengya, Tongzi; 15 Hanjiandian, Tongzi; 16 Guanyinqiao, Qijiang, S. Sichuan; 17 Shuanghe, Changning, SW Sichuan; 18 Yanjin, NE Yunnan; 19 Daguan, NE Yunnan; 20 Laokuangshan, Hongya, W. Sichuan; 21 Liangshan, Nanzheng, S. Shaanxi; 22 Xixiang, S. Shaanxi; 23 Zhenba, S. Shaanxi; 24 Bajaokou, Ziyang, S. Shaanxi; 25 Tangshan near Nanjing, Jiangsu; 26 Beigong, Jingxian, S. Anhui; 27 Tangjia, Yuqian, W. Zhejiang; 28 Xinkailing, Wuning, NW Jiangxi; 29 Shichengzi, Dajing, Gansu; 30 Xainza, Xizang (Tibet); 31 Mangjiu, Luxi, W. Yunnan: 32 Shahechang, Baoshan, W. Yunnan.

Churkin & Carter, *G. tamariscus linearis* Perner, *G.* ex gr. *tamariscus* Nicholson, *Climacograptus angustus* Perner, *C. bicaudatus* Chen & Lin and *C. normalis* Lapworth 0·20 m

(ACC764a) *Glyptograptus sinuatus* (Nicholson), *G. tamariscus linearis* Perner, *Climacograptus angustus* Perner, *C. wangjiawanensis* Mu & Lin, *Diplograptus modestus* Lapworth and *Rhaphidograptus minutus* Chen & Lin 0·04 m

11. Black argillaceous shale weathered brownish grey in colour, rich in graptolites including (ACC763d) *Glyptograptus persculptus* (Salter), *G. sinuatus* (Nicholson), *G.* ex gr. *tamariscus* Nicholson, *G. tamariscus linearis* Perner, *Diplograptus modestus* Lapworth, *Orthograptus guizhouensis* Chen & Lin, *Paraorthograptus innotatus* (Nicholson), *Climacograptus angustus* Perner, *C. normalis* Lapworth, *C. wangjiawanensis* Mu & Lin and *Rhaphidograptus minutus* Chen & Lin 0·16 m

(ACC763c) *Glyptograptus sinuatus* (Nicholson), *G. lunmaensis* Sun, *G. tamariscus linearis* Perner, *G. tamariscus magnus* Churkin & Carter, *Diplograptus* cf. *coremus* Chen & Lin, *Orthograptus angustifolius* Chen & Lin, *O. guizhouensis* Chen & Lin, *O. bellulus* Törnquist, *Climacograptus angustus* Perner and *C. wangjiawanensis* Mu & Lin 0·08 m

(ACC763b) *Glyptograptus sinuatus* (Nicholson), *G. lunmaensis* Sun, *G.* ex gr. *tamariscus* Nicholson, *G. tamariscus linearis* Perner, *G. tamariscus magnus* Churkin & Carter, *Diplograptus modestus* Lapworth, *Orthograptus angustifolius* Chen & Lin, *Paraorthograptus innotatus* (Nicholson), *P.* sp., *Climacograptus angustus* Perner and *C. normalis* Lapworth 0·06 m

(ACC763a) *Glyptograptus persculptus* (Salter), *G. sinuatus* (Nicholson), *G. lungmaensis* Sun, *G. tamariscus linearis* Perner, *G. tamariscus magnus* Churkin & Carter, *Diplograptus modestus* Lapworth, *Climacograptus angustus* Perner and *C. normalis* Lapworth 0·06 m

Upper Ordovician Wufeng Formation:

10. Bluish grey argillaceous calcareous silicolites weathered whitish-yellow and greyish-yellow, yielding abundant brachiopods and trilobites: (ACC762) *Leptaenopoma trifidum* Marek & Havlíček, *Kinnella kielanae* (Temple), *Dalmanella testudinaria* (Dalman), '*Paracraniops*' *patillis* Rong, *Cliftonia* cf. *oxoplecioides* Wright, *Hirnantia sagittifera* (M'Coy), *Draborthis* cf. *caelebs* Marek & Havlíček, *Aphanomena ultrix* (Marek & Havlíček), *Aegiromena* cf. *ultima* Marek & Havlíček and *Dalmanitina yichangensis* Lin, *D.* sp. 0·33 m

9. Black argillaceous shale and mudstone, yielding (ACC761) *Diplograptus bohemicus* (Marek) and *Paraothograptus typicus* Mu with a few brachiopods and cephalopods 0·26 m

8. Black shale intercalated with a few siliceous shale beds of the same colour, yielding: (ACC760) *Diplograptus bohemicus* (Marek), *D.* sp., *Glyptograptus* sp., *Climacograptus supernus* Elles & Wood and *Paraorthograptus* sp. 0·23 m

7. Black argillaceous shale with siliceous shale intercalation, yielding in the upper part (ACC759) *Dicellograptus ornatus* Elles & Wood, *Climacograptus supernus* Elles & Wood, *C. longicaudatus* Geh, C. sp., *Glyptograptus* sp., *Orthograptus truncatus* Lapworth and *Paraorthograptus uniformis* Mu & Li 0·42 m

Middle part (ACC758) *Tangyagraptus typicus* Mu, *Climacograptus supernus* Elles & Wood, *C. venustus* Hsu, *Amplexograptus suni* (Mu) and *Paraplegmatograptus* sp. 0·70 m

Lower part (ACC758a) *Dicellograptus szechuanensis* Mu, *D. ornatus* Elles & Wood, *Climacograptus supernus* Elles & Wood, *C.* sp., *Orthograptus truncatus* Lapworth, *Orthograptus maximus* Mu and *Amplexograptus suni* (Mu) 1·73 m

6. Black carbonaceous siliceous shale, yielding (ACC757) *Dicellograptus szechuanensis* Mu, *Amplexograptus disjunctus yangtzensis* Mu & Lin, *Pseudoclimacograptus* sp., *Orthograptus abbreviatus* Elles & Wood and *Parareteograptus sinensis* Mu 0·40 m

5. Black carbonaceous shale, yielding abundant graptolites: (ACC756) *Amplexograptus disjunctus yangtzeensis* Mu & Lin, *A. suni* (Mu), *Orthograptus* cf. *pauperatus* Elles & Wood and *Parareteograptus* sp. 0·43 m

4. Black carbonaceous shale intercalated with a few siliceous beds, yielding abundant graptolites (ACC755) *Leptograptus extremus modestus* Chen, *Dicellograptus* sp., *Climacograptus chiai* Mu, *Pseudoclimacograptus* spp., *Amplexograptus disjunctus yangtzeensis* Mu & Lin, *Orthograptus* cf. *maximus* Mu, *O. truncatus* Lapworth, *O.* cf. *pauperatus* Elles & Wood and *O.* sp. and inarticulate brachiopods 0·20 m

3. Dark grey to greyish green mudstone 0·12 m

Linhsiang Formation:

2. Dark yellow mudstone 0·05 m

1. Yellowish green to green argillaceous nodular limestone, yielding the trilobites (ACC754) *Hammatocnemis* sp. and *Microparia* sp. about 2·00 m

2. 'Baoshan' (the 'Treasure Hill') section, Huanghuachang, Yichang, W. Hubei (after Mu *et al.* 1984).

Lower Silurian Lungmachi Formation (basal part):

9. Black siliceous rock weathered greyish-yellow, yielding: (ACC744) *Parakidograptus acuminatus* (Nicholson), *Climacograptus normalis* Lapworth, *C. sinitzini* (Chaletzkaya) 0·10 m

8. Black carbonaceous shale, black siliceous shale weathered blackish grey, containing: (ACC743) *Glyptograptus persculptus* (Salter), *G. sinuatus* (Nicholson), *Climacograptus* sp. (cf. *normalis* Lapworth) 0·45 m

Upper Ordovician Wufeng Formation:

7. Black calcareous argillaceous siliceous mudstone weathered greyish-white to greyish-yellow, yielding abundant brachiopods, trilobites and other fossils, including (ACC742) *Hirnantia sagittifera* (M'Coy), *Kinnella kielanae* (Temple), *Aphanomena ultrix* (Marek & Havlíček), *Cliftonia* cf. *psittacina* (Wahlenberg), *Triplesia* sp., *Dalmanella testudinaria* (Dalman), *Aegiromena* cf. *ultima* (Marek & Havlíček), *Meristina crassa incipiens* (Williams) and *Dalmanitina yichangensis* Lin 0·10 m

5–6. Black argillaceous siliceous shale, weathered dark grey, yielding (ACC741) *Diplograptus bohemicus* (Marek) and a few brachiopods in the upper part 0·45 m

3–4. Black siliceous shale intercalated with argillaceous shale, containing (ACC740) *Dicellograptus ornatus* Elles & Wood, *D.* sp., *Glyptograptus* sp., *Climacograptus supernus* Elles & Wood, *C. hastatus* Hall, *C.* sp. and *Paraorthograptus uniformis* Mu & Li 0·51 m

2. Black shale intercalated with black siliceous shale, yielding (ACC739) *Diceratograptus mirus* Mu, *D. ornatus brevispinus* Chen, *Glyptograptus* sp., *Climacograptus hastatus* Hall 0·20 m

1. Black shale with a few siliceous shale intercalations, rich in graptolites including (ACC737) *Tangyagraptus uniformis* Mu, *Dicellograptus ornatus* Elles & Wood, *D. ornatus brevispinus* Chen, *Glyptograptus* sp., *Climacograptus supernus* Elles & Wood, *C. supernus longus* Geh, *C. tumidus* Geh, *Amplexograptus suni* (Mu), *Orthograptus abbreviatus* Elles & Wood, *Yinograptus disjunctus* (Yin & Mu), *Y. brevispinus* Mu, *Paraplegmatograptus connectus* Mu 0·15 m

Black shale with siliceous shale intercalation, yielding abundant graptolites, including (ACC737a) *Tangyagraptus typicus* Mu, *T. uniformis* Mu, *T.* sp., *Climacograptus supernus* Elles & Wood, *C. supernus longus* Geh, *Orthograptus truncatus* Lapworth, *Glyptograptus* sp., *Amplexograptus suni* (Mu), *Yinograptus disjunctus* (Yin & Mu), *Y. grandis* Mu, *Paraplegmatograptus* sp. [ ]

3. Renhuai section (after Geng Liang-yu *et al.* 1984).

Lower Silurian Lungmachi Formation (basal part):

Greyish-black silty, carbonaceous shale (0·05 m thick in single bed), cream-coloured sandy shale (in basal part), yielding an abundant graptolite fauna of *Glyptograptus kaochiapienensis* Hsu, *G.* cf. *lungmaensis* Sun and *Orthograptus* sp. etc. associated with some brachiopods 1·8 m

Upper Ordovician Wufeng Formation:

2. Kuanginchiao bed, including the following units:

c. dark grey thick-bedded bioclastic limestone in upper part (ADR557-3) with numerous solitary corals such as *Brachylasma* sp., *Crassilasma* sp. and *Dansiphyllum*? sp. 1·14 m

b. Dark greyish thin-bedded bioclastic limestone in the middle part (ADR557-2) including *Hirnantia sagittifera* (M'Coy), *Dalmanella testudinaria* (Dalman), *Aphanomena ultrix* Marek & Havlíček, *Dalmanitina* sp., *Modiolopsis* sp., rugose corals, and the chitinozoan *Conochitina* cf. sp. A of Achab 0·29 m

a. Dark greyish medium-bedded limestone in lower part (ADR557-1) with the monotomous chitinozoan *Conochitina* cf. sp. A of Achab 0·67 m

1. Greyish-black carbonaceous shale with a minor quantity of clayey shale in the upper part, dark greyish dolomitic limestone in the lower part and 4 cm greyish black carbonaceous shale in basal part, yielding abundant graptolites such as *Climacograptus hastatus* Hall, *C.* sp., *Paraorthograptus typicus* Mu, *P.* sp., *Dicellograptus ornatus* Elles & Wood, *D. tenuisculus* Mu *et al.*, *D. szechuanensis* Mu and *Pleurograptus lui* Mu 4·1 m

4. The Nanzheng Formation of Liangshan, Nanzheng county, S. Shaanxi, was considered to be basal Silurian for a long time. However Zhu *et al.* (1986) have revised this to a late Ordovician age. According to their detailed work, the Nanzheng Formation is the equivalent of the Wufeng Formation and indicates a mixed biofacies. The Liangshan Ordovician–Silurian boundary section, Nanzheng, measured by them may be summarized as follows:

Lower Silurian Lungmachi Formation (basal part):

11. Brownish grey shales with *Climacograptus angustus* (Perner), *Diplograptus uniformis* Li, *Glyptograptus lungmaensis* Sun, *G. tamariscus distans* Packham, *G. tamariscus linearis* Perner 0·5 m

10. Brownish grey and pinkish shale with a few cephalopods and brachiopods (NZ10) and *Climacograptus normalis* Lapworth, *C. miserabilis* Elles & Wood, *C. angustus* (Perner), *Diplograptus* ex gr. *modestus* Lapworth, *D. uniformis* Li, *Glyptograptus lungmaensis* Sun 0·27–0·32 m

Upper Ordovician Nanzheng Formation:

9. Brownish-yellow calcareous shale rich in (NZ9) *Climacograptus angustus* (Perner), *Orthograptus* sp., *Glyptograptus* sp., *Platycoryphe sinensis* (Lu), *Dalmanitina* sp.; the bivalve *Deceptrix* sp. and some compressed cephalopods 0·17–0·22 m

8. Brownish-grey medium-bedded argillaceous limestone with (NZ8) *Diplograptus* cf. *bohemicus* (Marek), *Orthograptus* sp., *Climacograptus* sp., *Pleurorthoceras shanchongense* Zou, *P. jingxianense* Zou, *P. slendertubulatum* Zou, *P.* cf. *clarksvillense* (Foerste), *Michelinoceras* sp., *Aegiria*? sp., *Platycoryphe sinensis* (Lu) and *Dalmanitina nanchengensis* Lu 0·74 m

7. Brownish argillaceous limestone, containing (NZ7) *Dalmanitina nanchenensis* Lu, *Platycoryphe sinensis* (Lu), the gastropod *Rhaphistomina*? sp., and brachiopod fragments 0·46 m

6. Brownish to light grey, coarse quartzitic sandstone 0·83 m

5. Light brown shale intercalated with sandstone containing bivalve fragments in the top part (NZ6) 2·30 m

4. Greyish shale containing a few graptolites (NZ5) including *Climacograptus* sp. 0·25 m

3. Grey clayey and aluminal shale rich in fossils (NZ4) with *Orthograptus maximus* Mu, *O.* cf. *abbreviatus* Elles & Wood, *Climacograptus normalis* Lapworth, *Diplograptus* sp., *Parareteograptus* sp., *Dictyonema* sp., *Orbiculoidea*, *Euklesdenella*, the bryozoans *Stictopora*, *Hallopora* and *Escharopora*; *Conularia* and *Metoconularia* (?) *proteica* (Barrande) 0·28 m

2. Light grey siliceous shale containing (NZ2) *Orthograptus maximus* Mu, *Climacograptus angustus* (Perner) in the lower part 0·15 m

1. Light grey and brownish siltstone and shale 0·5 m

Linhsiang Formation:

Light green and brownish argillaceous limestone, with *Nankinolithus* sp. and *Protopanderodus insculptus* (Branson & Mehl) in the upper (NZ2) and *Paraceraurus* cf. *longisulcatus* Lu in the lower (NZ1) 1·10 m

5. Gaojiawan section, Xixiang, S. Shaanxi. A most detailed Ordovician–Silurian section was measured by Yu *et al.* (1986) as follows:

Lower Silurian Lungmachi Formation:

10. Black siliceous and carbonaceous shale containing (XF162–155) *Orthograptus vesiculosus* (Nicholson), *Climacograptus transgrediens* Waern and *C. medius* Törnquist. 2·77 m

9. Black siliceous shale interbedded with carbonaceous shale rich in graptolites (XF154–135) with *Parakidograptus acuminatus* (Nicholson), *Akidograptus ascensus* Davies, *A. xixiangensis* Yu, Fang & Zhang, *A. parallelus* Li & Jiao, *Climacograptus sinitzini* (Chaletzkaya) and *Orthograptus lonchoformis* Chen & Lin 4·63 m

8. Black siliceous shale intercalated with black carbonaceous shale rich in graptolites (XF134–125) with *Glyptograptus persculptus* Salter, *G. persculptus–sinuatus* transient, *G. tamariscus* (Nicholson), *G. lungmaensis* Sun, *G. zhui* Yang, *Climacograptus normalis* Lapworth, *Orthograptus lonchoformis* Chen & Lin, *Akidograptus ascensus* Davies and *A. xixiangensis* Yu, Fang & Zhang 0·89 m

Upper Ordovician Wufeng Formation:

7. Black siliceous shale weathered purplish brown in colour, containing (XF124–118) *Diplograptus bohemicus* (Marek), *D. orientalis* Mu, *Climacograptus normalis* Lapworth, *Glyptograptus* sp. 0·64 m

6. Greyish to pale siltstone and quartzitic sandstone containing (XF117–115) *Dalmanitina wuningensis* Liu, *Leonaspis* (*Eoleonaspis*) *olinini* (Troedsson), *Hirnantia sagittifera* (M'Coy), *Kinnella kielanae* (Temple) 0·22 m

5. Black siliceous and carbonaceous shale rich in graptolites (XF114–112) with *Paraorthograptus uniformis* Mu & Li, *Orthograptus truncatus* Lapworth, *Climacograptus hastatus* Hall, *Paraplegmatograptus* sp. and *Dicellograptus* sp. 0·26 m

4. Black carbonaceous shale and siliceous shale containing graptolites (XF111–110) *Paraorthograptus typicus* Mu, *Climacograptus supernus* Elles & Wood, *C. hastatus* Hall, *Paraplegmatograptus* sp., *Dicellograptus graciliramosus* Yin & Mu 0·17 m

3. Black shale weathered brown, containing (XF109–107) *Tangyagraptus typicus* Mu, *Paraorthograptus typicus* Mu, *Climacograptus hastatus* Hall, *C. venustus* Hsu, *Amplexograptus suni* (Mu), *Dicellograptus ornatus* Elles & Wood, *Yinograptus disjunctus* (Yin & Mu), *Parareteograptus* sp. 0·33 m

2. Dark grey shale with (XF106–104) *Dicellograptus szechuanensis* Mu, *D. excavatus* Mu, *Pleurograptus lui* Mu, *Climacograptus supernus* Elles & Wood, *Parareteograptus sinensis* Mu, *Orthoreteograptus denticulatus* Mu 0·42 m
1. Dark grey to black shale, containing (XF103–101) *Pleurograptus lui* Mu, *Dicellograptus elegans* Carruthers, *Climacograptus supernus* Elles & Wood, *Pseudoclimacograptus* sp., *Glyptograptus* sp., *Parareteograptus sinensis* Mu, *Orthoreteograptus denticulatus* Mu 0·44 m

Jiancaogou Formation:
Grey and yellowish green mudstone with *Nankinolithus*, etc.

In the section listed above, unit 1 is the *Pleurograptus lui* Zone which is equivalent to the *Amplexograptus disjunctus yangtzensis* Zone (W$_1$). Unit 2 is the *Dicellograptus szechuangensis* Zone (W$_2$) and unit 3 is the *Tangyagraptus typicus* Zone (W$_3$). Unit 4 is the equivalent of the *Diceratograptus mirus* Zone (W$_4$) but *D. mirus* itself has not been found. Unit 5 is the *Paraorthograptus uniformis* Zone (W$_5$), unit 6 is the *Hirnantia–Dalmanitina* bed (HD) and unit 7 is the *Diplograptus bohemicus* Zone (W$_6$). Unit 8 is the *Glyptograptus persculptus* Zone (L$_1$) characterized by the occurrence of *G. persculptus*, *G. persculptus–sinuatus* transient, *G. zhui* and *G. lungmaensis*. It is noteworthy that *Akidograptus ascensus* first appears in the lower part of this zone and *A. xixiangensis* appears in the upper part. Unit 9 is the *Parakidograptus acuminatus* Zone (L$_2$) characterized by the incoming of *P. acuminatus* and *Climacograptus sinitzini* in association with *A. ascensus* and *A. xixiangensis*. Unit 10 is the *Orthograptus vesiculosus* Zone (L$_3$) characterized by the incoming of *O. vesiculosus*.

6. Bajaokou Ordovician–Silurian boundary section, Ziyang county, S. Shaanxi. The Lower Silurian Banjuguan Formation and the Upper Ordovician Bajaokou Formation are all in graptolite facies, without shelly beds. They are composed of dark grey to black carbonaceous and siliceous slate and rich in graptolites, which were deposited in deep water on the south slope of the East Qinling trough and on the north margin of the Yangtze platform. The thickness of the basal Silurian is much greater than that of the uppermost Ordovician. The section measured by Fu and others may be outlined as follows.

Lower Silurian Banjiuguan Formation (basal part). Black carbonaceous and siliceous slate:
L$_3$ *Orthograptus vesiculosus* Zone with *O. vesiculosus*, *Neodicellograptus*, *Rhaphidograptus*, and *Atavograptus* 27·4 m
L$_2$ *Parakidograptus acuminatus* Zone with *P. acuminatus* and *Climacograptus sinitzini* (F14) 20·8 m
L$_1$ *Glyptograptus persculptus–sinuatus* transient zone 10·5 m
 4. *G. persculptus–sinuatus* transient, and *G. tamariscus* (F13)
 3. *Akidograptus ascensus*, *Climacograptus miserabilis*, *Orthograptus*, and *Atavograptus* (F12)
 2. *Glyptograptus* cf. *persculptus*, *Orthograptus lonchoformis* and *Diplograptus* cf. *modestus* (F11)
 1. *G.* cf. *persculptus*, *G. sinuatus*, *G. gracilis*, *Diplograptus modestus*, *Climacograptus normalis*, and *C. miserabilis* (F10)

Upper Ordovician Bajaokou Formation (upper part). Dark grey to black carbonaceous and siliceous slate:
W$_6^2$ *Diplograptus* spp., *Climacograptus* sp., *Orthograptus* sp. (F9, F8) 2 m
W$_6^1$ *Climacograptus extraordinarius*, *Diplograptus* spp. (F7, F6) 1·5 m
W$_5$ *Paraorthograptus uniformis* (F4) 1·2 m
W$_4$ *Diceratograptus mirus* (F3) 0·6 m

7. Tangshan Ordovician–Silurian boundary section near Nanjing (Jiao & Zhang 1984).

Lower Silurian Kaochiapien Formation (basal part):
10. Greyish and yellowish shale with chert (ND8), containing *Glyptograptus caudatus* Ge, *Climacograptus normalis* Lapworth, and *Orthograptus* sp. 0.30 m
9. Variegated siliceous shale with (ND7) *Glyptograptus lungmaensis* Sun, *Orthograptus* sp. and *Akidograptus*? sp. 0·40 m
8. Purple siliceous shale rich in graptolites (ND6) with *Diplograptus* sp., *Glyptograptus* sp. and *Climacograptus* sp. 0·02 m

Upper Ordovician Wufeng Formation:

7. Kuanyinchiao bed: greyish siliceous mudstone rich in shelly fossils (ND5) with *Dalmanitina yichangensis* Lin, *Leonaspis sinensis* Chang, *Platycoryphe* sp., *Paromalomena polonica* (Temple), *Aegiromena ultima* Marek & Havlíček, *Triplesia*? sp., *Holopea*? sp., *Loxonema* sp., *Nuculoidea* sp. and *Hyolithes*? 0·19 m

6. Black sandy shale (ND4), containing *Diplograptus* cf. *bohemicus* (Marek) and *Climacograptus extraordinarius* (So6) 0·28 m

5. Variegated calcareous mudstone 0·09 m

4. Purple greyish siliceous shale with graptolites (ND3) *Diplograptus* sp. and *Climacograptus* sp. 0·09 m

3. Brownish yellow shale (ND2) with the brachiopod *Manosia* sp., the gastropod *Planetochidea* and trilobite and crinoid fragments. 0·30 m

2. Grey siliceous pale-weathered shale 0·45 m

1. Black siliceous shale with (ND1) *Dicellograptus* sp. and *Climacograptus supernus* Elles & Wood 0·83 m

8. Xainze area, Northern Xizang (Tibet) (after Mu & Ni, 1983).

Lower Silurian Dewukaxia Formation (basal part):

Black graptolitic shale with *Climacograptus normalis* Lapworth, *C. miserabilis* Elles & Wood, *C. xainzaensis* Mu & Ni, *Glyptograptus elegantulus* Mu & Ni, *G. nanus* Mu & Ni, *G. asthenus* Mu & Ni, *Diplograptus lacertosus* Mu & Ni, *D. spanis* Mu & Ni and *D. temalaensis* (Jones).

Upper Ordovician Xainza Formation:

Grey argillaceous limestone with *Hirnantia*, *Kinnella*, *Cliftonia*, *Paromalomena*, *Hindella*, *Aphanomena* and dalmanitid trilobite 8·82 m

Greyish-yellow shale with *Glyptograptus asthenus* Mu & Ni, *G. daedalus* Mu & Ni, *G. elegantulus* Mu & Ni, *G. nanus* Mu & Ni, *Diplograptus bohemicus* (Marek), *D. charis* Mu & Ni, *D. flustrianus* Mu & Ni, *D. maturatus* Mu & Ni, *D. ojsuensis* (Koren & Mikhaylova), *D. orientalis* Mu *et al.*, *D. spanis* Mu & Ni, *D. viriosus* Mu & Ni, *Climacograptus* cf. *extraordinarius* (Sobolevskaya), *C. miserabilis* Elles & Wood, *C. normalis* Lapworth, *C. xainzaensis* Mu & Ni, *C. xizangensis* Mu & Ni and *Orthograptus* sp. 5·27 m

Upper Ordovician Gangmusang Formation:

Limestone with shelly fauna.

9. Mangjiu section of Luxi (after Ni *et al.*, 1983).

Lower Silurian Lower Jenhochiao Formation (basal part):

4. Black shale with *Climacograptus normalis* Lapworth, *C. miserabilis* Elles & Wood, *C. trifilis lubricus* Chen & Lin, *Akidograptus ascensus* Davies, *Orthograptus guizhouensis* Chen & Lin, *Diplograptus bifurcus* Mu *et al.*, etc. 4·1 m

3. Sandy mudstone with *Climacograptus normalis* Lapworth and *C.* sp. c. 0·5 m

Upper Ordovician Wanyaoshu Formation (top part):

2. Greyish-white mudstone with *Hirnantia sagittifera* (M'Coy), *Hindella crassa incipiens* (Williams), *Coolinia* cf. *dalmani* Bergström, *Plectothyrella* cf. *crassicosta* (Dalman), *Paromalomena polonica* (Temple), *Aphanomena ultrix* Marek & Havlíček and *Dalmanitina* sp. c. 2 m

1. Black shale, containing *Climacograptus latus* Elles & Wood, *C. angustus* Perner and *Orthograptus maximus* Mu.

10. The Ordovician–Silurian boundary strata are well developed at the locality of Shahechang, about 15 km NW of Baoshan, Yunnan, where a number of graptolites were collected from the uppermost Ordovician by Ni Yu-nan, Cai Cong-yang, Chen Ting-en, Li Guo-hua, and Wang Ju-de. The stratigraphical sequence is as follows (in descending order):

Lower Silurian Lower Jenhochiao Formation (basal part):

3. Upper part: Black siliceous shale with *Pristiograptus* sp. and *Climacograptus* sp.

Lower part: Greyish white sandy shale with *Climacograptus normalis* Lapworth, *C. xainzaensis* Mu & Ni and *Glyptograptus* sp. (ex gr. *persculptus*) in the basal 2 m.

Upper Ordovician:

2. Greyish black sandy shale, rich in graptolites, the top part with *Diplograptus bohemicus* (Marek), *Diplograptus ojsuensis* (Koren & Mikhaylova), *Climacograptus normalis* Lapworth (ACJ196), *Cli-*

macograptus cf. *normalis* Lapworth, *C. xainzaensis* Mu & Ni, *C. extraordinarius* (Sobolevskaya), *Diplograptus* cf. *orientalis* Mu *et al.*, *D. yunnanensis* Ni (ACJ195). The middle part yields *Glyptograptus daedalus* Mu & Ni and *Climacograptus extraordinaris* (Sobolevskaya) (ACJ194); and the basal part *Glyptograptus* cf. *elegantulus* Mu & Ni, *G. daedalus* Mu & Ni, *Diplograptus maturatus* Mu & Ni, *D. ojsuensis* (Koren & Mikhailova) and *D. temalaensis* (Jones) (ACJ193).

1. Yellow argillaceous limestone with *Nankinolithus*? sp., *Cyclopyge* sp., etc.

Analysis of the boundary sections

The strata across the Ordovician–Silurian boundary in China fall into different biofacies types as follows.

1. Where the graptolitic *Glyptograptus persculptus* Zone (L_1) lies upon the graptolitic *Diplograptus bohemicus* Zone (W_6) without intervening shelly beds, as in the Bajaokou section, Ziyang, S. Shaanxi.
2. Where the graptolitic *Glyptograptus persculptus* Zone or its equivalents (L_1) lies upon the graptolitic *Diplograptus bohemicus* Zone (W_6) with a shelly bed below, as in the Xixiang section, Xixiang, S Shaanxi; the Ganxi section, Yanhe, NE Guizhou; and the Shahechang section, Baoshan, W Yunnan.
3. Where the graptolitic facies with the *Glyptograptus persculptus* Zone or its equivalents (L_1) lies upon shelly *Hirnantia–Dalmanitina* beds (HD) with a graptolitic facies below, as at the Wangjiawan, Huanghuachang, Fenxiang and Tangya Sections, all in Yichang, W Hubei; the Sintan section, Zigui, W Hubei; the Shuanghezhen section, Changning, SW Sichuan; the Guanyiqiao section, Qijiang, S Sichuan; the Xiushan section, SE Sichuan; the Songtao section, NE Guizhou; the Hanjiadian and Liangfengya sections, Tongzi, N Guizhou; the Renhuai and Bijie sections, NW Guizhou; the Yanjin and Daguan sections, NE Yunnan; the Luxi section, W Yunnan; and the Xainza sections of Xizang (Tibet).
4. Where the graptolitic facies with *Glyptograptus persculptus* or its equivalents (L_1) lies upon a mixed facies with graptolitic facies below, such as in the Honghuayuan section, Tongsi, N Guizhou; the Liangshan section, Nanzheng, S Shaanxi; the Xinkailing section, Wuning, NW Jiangxi; the Shanchong section, Jingxian, S Anhui; and the Tangjia section, Yuqiau, W Zhejiang.
5. Where the shelly Wulipo bed with an '*Eospirigerina*' fauna lies upon the shelly *Hirnantia–Dalmanitina* bed with graptolitic facies below, as at Donggongsi, Zunyi, in N Guizhou.

Strata of the first type are only known in the transitional belt between the Yangtze basin and the East Qinling trough to the north, whereas the last type is only known in the southern marginal belt of the Yangtze basin. The Ordovician–Silurian boundary sections of the second and fourth types are important for the correlation of the *Diplograptus bohemicus* Zone (W_6) and the *Hirnantia–Dalmanitina* fauna (HD). The Ordovician–Silurian boundary sections of the third type are most common and widespread in the Yangtze region. The Wufengian (Ashgill) Yangtze sea was bounded by surrounding lands and swells and became a semi-enclosed sea under aerobic conditions, but the surface water above the anoxic layer was oxygenated. The strata of the third type are rich in organic matter and graptolites flourished.

The diversity of the Wufeng graptolitic fauna increases upwards stratigraphically from the *Amplexograptus disjunctus yangtzeensis* Zone (W_1) to the *Tangyagraptus typicus* Zone (W_3). More than twenty genera occur in the *Dicellograptus szechuanensis* Zone (W_2), apart from the dendroids. The decline of graptolite diversity took place from the *Diceratograptus mirus* Zone (W_4) to the *Diplograptus bohemicus* Zone (W_6) (Table 2). At the end of the Ordovician, all the axonolipous graptoloids were nearly extinct except for a few *Dicellograptus* which remained in China. In contrast, the Wufengian benthic shelly fauna increased in diversity. The well-known, cosmopolitan *Hirnantia* fauna first appeared in the equivalents of the *Diceratograptus mirus* Zone (W_4) with 7 genera, and increased gradually to 23 genera in the uppermost Ordovician *Hirnantia–Dalmanitina* bed (Table 3). The sea level was lowered in late Ordovician due to the formation of the ice cap in North Africa. In the late Wufengian W_4–W_6, a shallow and better aerated environment occurred due to ventilation of sea waters. The maximum glaciation was

Table 2 Stratigraphical range of graptolite genera in the Wufeng Formation

	W_1	W_2	W_3	W_4	W_5	W_6
Leptograptus	+	+	+	—	—	—
Pleurograptus	+	+	+	—	—	—
Dicellograptus	+	+	+	+	+	+
Diceratograptus	—	—	—	+	—	+
Dicranograptus	—	+	—	—	—	—
Tangyagraptus	—	—	+	—	—	—
Glyptograptus	+	+	+	+	+	+
Amplexograptus	+	+	+	+	+	+
Climacograptus	+	+	+	+	+	+
Pseudoclimacograptus	+	+	—	—	—	—
Diplograptus	—	+	+	+	+	+
Orthograptus	+	+	+	+	+	—
Paraorthograptus	—	+	+	+	+	+
Parareteograptus	+	+	+	—	—	—
Orthoreteograptus	+	+	—	—	—	—
Sinoreteograptus	+	—	—	—	—	—
Neurograptus	+	+	—	—	—	—
Nymphograptus	+	+	—	—	—	—
Arachniograptus	—	+	+	—	—	—
Phormograptus	—	+	+	—	—	—
Plegmatograptus	+	—	—	—	—	—
Paraplegmatograptus	—	+	+	+	+	+
Yinograptus	—	+	+	—	—	—
Yangzigraptus	—	—	+	+	—	—

Table 3 Stratigraphical range of brachiopod genera in the Upper Wufeng Formation

	W_4	W_5	W_6^1	W_6^2
Paracraniops	+	+	+	+
Dalmanella	+	+	+	+
Paromalomena	+	+	+	+
Leptaena	+	+	+	+
Aphanomena	+	+	+	+
Coolinia	+	+	+	+
Hindella	+	+	+	+
Trematis	—	+	—	—
Hirnantia	—	+	+	+
Cliftonia	—	+	+	+
Plectothyrella	—	+	+	+
Dorytreta	—	+	+	+
Philhedra	—	—	+	+
Philhedrella	—	—	+	+
Acanthocrania	—	—	+	+
Kinnella	—	—	+	+
Draborthis	—	—	+	+
Mirorthis	—	—	+	+
Aegiromena	—	—	+	+
Leptaenopoma	—	—	+	+
Toxorthis	—	—	—	+
Dysprosorthis	—	—	—	+
Trucizetina	—	—	—	+
Onychoplecia	—	—	—	+

reached at the end of the Ordovician (W_6) and the whole Yangtze basin became a nearly normal shallow sea in which the *Hirnantia–Dalmanitina* fauna flourished.

At the beginning of the Silurian a new graptolite fauna occurred, notably with monograptids and typical Silurian diplograptids such as the *Diplograptus* cf. *modestus* and *Glyptograptus* cf. *tamariscus* groups during the *Glyptograptus persculptus* Zone (L_1) time interval. A new brachiopod fauna, known as the '*Eospirigerina*' fauna, appeared above the *Hirnantia* fauna in the nearshore region. The rapid change in biofacies and faunal composition is due to the rising of sea level caused by rapid melting of the ice cap.

Correlation of the Ordovician–Silurian boundary sections

All the Ordovician–Silurian boundary sections may be easily correlated in China by the standard of the Wufengian–Lungmachian graptolite zones and the *Hirnantia–Dalmanitina* bed. In order to define the Ordovician–Silurian boundary throughout the world, a precise correlation of the *Diplograptus bohemicus*, *Glyptograptus persculptus* and *Parakidograptus acuminatus* Zones with shelly faunas is necessary. Thus, the subdivision and correlation of the *Diplograptus bohemicus* Zone with the *Hirnantia–Dalmanitina* bed is of great importance.

In the Yichang sections, Western Hubei, the uppermost *Hirnantia–Dalmanitina* bed is underlain by the *Diplograptus bohemicus* Zone (W_6) and overlain by the *Glyptograptus persculptus* Zone (L_1), whereas in the Xixiang section, S. Shaanxi, the *Hirnantia–Dalmanitina* bed is underlain by the *Paraorthograptus uniformis* Zone (W_5) and overlain by the *Diplograptus bohemicus* Zone (W_6), which is succeeded by the *Glyptograptus persculptus* Zone (L_1). Therefore, the *D. bohemicus* Zone of Yichang is equivalent to the lower part of the *D. bohemicus* Zone (W_6^1), and the *D. bohemicus* Zone of Xixiang is equivalent to the upper part of the *D. bohemicus* Zone (W_6^2). Thus the *Hirnantia–Dalmanitina* bed of Yichang is the equivalent of the upper part of the *D. bohemicus* Zone (W_6^2), and that of Xixiang is the equivalent of the lower part of the *D. bohemicus* Zone (W_6^1). *Climacograptus extraordinarius* and *Diplograptus orientalis* usually occur in the lower part of the *D. bohemicus* Zone (W_6^1).

The *Glyptograptus persculptus* Zone (L_1) is marked by the incoming of *Glyptograptus persculptus*, *G. sinuatus*, *G. lungmaensis*, *G. gracilis*, *Diplograptus modestus*, *Akidograptus ascensus* and monograptids. It represents the beginning of a new developmental stage of graptolite faunas, the fifth (or monograptid) fauna as defined by the writer (Mu 1984). Thus the base of the *G. persculptus* Zone should be considered an important stratigraphical boundary, that between the Ordovician and Silurian.

It is noteworthy that the *Akidograptus ascensus* Zone, directly overlying the *Hirnantia–Dalmanitina* beds of Europe, is usually regarded as the equivalent of *Parakidograptus acuminatus* by some foreign colleagues. For defining the Ordovician–Silurian boundary the correlation of the *Akidograptus ascensus* Zone with the *Glyptograptus persculptus* Zone and the boundary between the *Glyptograptus persculptus* Zone and the *Parakidograptus acuminatus* Zone must be clarified.

The *Parakidograptus acuminatus* Zone (L_2) is marked by the incoming of *P. acuminatus* in association with *Climocograptus sinitzini* which also characterizes the *P. acuminatus* Zone. *Akidograptus ascensus* itself first appeared in the *persculptus* Zone (L_1), much earlier than *P. acuminatus*, although the two forms may be present together in the *P. acuminatus* Zone (L_2), whereas *P. acuminatus* is confined to the *P. acuminatus* Zone. Yu and his colleagues are of the opinion that *Parakidograptus acuminatus* is directly derived from *Akidograptus ascensus* and a transitional form *Akidograptus xixiangensis* Yu *et al.* was described and illustrated from the basal Lungmachi formation of Xixiang, S. Shaanxi. *A. xixiangensis* appears higher than *A. ascensus* and lower than *P. acuminatus.* It posseses akidograptid thecae in the proximal portion of the rhabdosome and parakidograptid thecae in the distal portion. A similar form *Akidograptus giganteus* was described by Yang (1964) from the basal Silurian of W. Zhejiang. Li & Ge (1981) and Fu (1983) tried to propose a new genus for these transitional forms between *Akidograptus* and *Parakidograptus.*

It is clear that the *Akidograptus ascensus* Zone of Europe may be correlated with the *Glyptograptus persculptus* Zone in China. This view was confirmed by the works of Nilsson (1984) in Sweden, and Storch (1982) in Bohemia. The same is true, in my view, for the Mirny Creek section, northeast USSR, studied by Koren *et al.* (1983). The Mirny Creek Ordovician–Silurian boundary section of mixed biofacies measured by Koren and her colleagues may be outlined mainly by graptolites as follows:

Members 65 and 66 *Paraorthograptus pacificus* Zone

Members 67 and 68 *Climacograptus extraordinarius* Zone with *Hirnantia–Dalmanitina* fauna

Members 69 to 72 *Diplograptus bohemicus* Zone (='*persculptus*' Zone) with *Hirnantia–Dalmanitina* fauna

Members 73 and 74 *Akidograptus ascensus* Zone, incoming of *Diplograptus* of *modestus* group, *Glyptograptus* of the *tamariscus* group and *Akidograptus ascensus.*

Members 75 to basal part of member 78 *Parakidograptus acuminatus* Zone, incoming of *P. acuminatus* and *Climacograptus sinitzini..*

Member 78 *Orthograptus vesiculosus* Zone, incoming of *Orthograptus vesiculosus.*

It is obvious that the *Paraorthograptus pacificus* Zone (65–66) corresponds to the *Paraorthograptus uniformis* Zone (W_5), that the *Climacograptus extraordinarius* Zone (67–68) corresponds to the lower part of the *Diplograptus bohemicus* Zone (W_6^1), and the *Diplograptus bohemicus* Zone (='*persculptus*' Zone, 69–72) corresponds to the upper part of the *Diplograptus bohemicus* Zone (W_6^2). The lower part of the '*acuminatus–ascensus* Zone' (members 73–74) of Koren and others is equivalent to the *Akidograptus ascensus* Zone of Europe, and corresponds to the *Glyptograptus persculptus* Zone (L_1) of China, whereas the upper part of the '*acuminatus–ascensus* Zone' (75–basal 78) is the *Parakidograptus acuminatus* Zone, corresponding to the *Parakidograptus acuminatus* Zone (L_1) of China and Europe.

I am convinced that the *Akidograptus ascensus* Zone of the European continent is equivalent to the *Glyptograptus persculptus* Zone of Britain and Denmark. The *Parakidograptus acuminatus* Zone and the *Glyptograptus persculptus* Zone of the Dob's Linn section of Britain correspond to the *P. acuminatus* Zone (L_2) and *G. persculptus* Zone (L_1) of China respectively. The *C. extraordinarius* band of the Dob's Linn section falls within the lower part of the *Diplograptus bohemicus* Zone (W_6^1), and the blind dalmanitid band of Dob's Linn possibly falls within the upper part of the *D. bohemicus* Zone (W_6^2). It seems to me that the *G. persculptus* Zone of Dob's Linn as well as elsewhere represents the beginning of the Silurian transgression due to the rapid melting of the ice-cap in North Africa.

Conclusions

1. The Ordovician–Silurian boundary sections are widely distributed in China. Many Ordovician–Silurian boundary sections have been defined in the Yangtze platform of the Central China Region.

2. The graptolite sequence of the upper Ordovician (Wufengian W_1–W_6) and the Lower Silurian (Lungmachian L_1–L_7) affords a valuable standard for correlation. The position of the *Hirnantia–Dalmanitina* bed is confined to W_4–W_6. The *Diplograptus bohemicus* Zone (W_6) is the highest level reached by the well-known and cosmopolitan *Hirnantia* fauna.

3. By this standard all the Ordovician–Silurian boundary sections may be easily correlated in China and even outside China.

4. The *acuminatus* Zone is marked by the incoming of *Parakidograptus acuminatus.* The underlying *Akidograptus ascensus* Zone of Europe is equivalent to the *Glyptograptus persculptus* Zone, which is the beginning of the Silurian transgression due to the rapid melting of the ice-cap in north Africa. The *G. persculptus* Zone was also the beginning of the monograptid fauna stage in the history of the development of the graptolite faunas. It is reasonable to place the Ordovician–Silurian boundary between the *G. persculptus* Zone (L_1) or '*Eospirigerina*' bed and the *D. bohemicus* Zone (W_6) or the *Hirnantia–Dalmanitina* bed (HD).

5. The *C. extraordinarius* Zone of the north-east USSR or the *C. extraordinarius* band of Dob's Linn, Scotland, correspond to the lower part of the *D. bohemicus* Zone (W_6^1). The '*G.*

persculptus' (= *D. bohemicus*) Zone of the north-east USSR corresponds to the upper part of the *D. bohemicus* Zone (W_6^2) of China.

6. Many kinds of fossils have been found in the Ordovician–Silurian boundary sections such as graptolites, brachiopods, trilobites, ostracods, corals, bivalves, cephalopods, gastropods, bryozoa, crinoids, conularia, conodonts, chitinozoa, and so on. The increasing number of finds of conodonts is of great importance for correlation with the Anticosti section of Canada. At present, the correlation with Anticosti is difficult. Unfortunately there are many weak points in the Dob's Linn section, and it is difficult to use as an international Ordovician–Silurian boundary stratotype.

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The Ordovician–Silurian boundary beds of the north-east USSR

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Synopsis

Graptolites of the *supernus*, *extraordinarius*, *persculptus*, *acuminatus* and *ascensus* Zones are present in sections in the north-east USSR, with the best section at Mirny Creek. Brachiopod and coral faunas also occur with the *Tcherskidium* and *Holorhynchus* beds in the *supernus* Zone and the *Hirnantia*? beds present in the *persculptus* Zone, both within the Tirekhtyakh Horizon. The succeeding *acuminatus* and *ascensus* Zone graptolites are developed in the Chalmak Horizon, which also bears a sparse shelly fauna.

Introduction

The late Ordovician and early Silurian boundary beds in the north-east USSR crop out on the Omulev Uplift in the upper Kolyma Basin. They are built up by terrigenous-carbonate and terrigenous deposits which are variable in composition and contain a mixed shelly-graptolite fauna. The rocks are exposed on limbs of extensive anticlines and show either a monoclinal succession, such as at Mirny Creek, Neznakomka River and Drevnyaya River, or represent large fragments of sections among complex faulted sequences, such as at the Ina River. The Upper Ashgill and Lower Llandovery deposits include the *supernus*, *extraordinarius*, *persculptus*, *acuminatus* and *ascensus* graptolite Zones and have a total thickness of about 300 m (Fig. 1). This part of the section is designated the Tirekhtyakh and Chalmak horizons. The lower part of the Tirekhtyakh horizon (the *supernus* Zone) (Fig. 2) shows a diversity of facies from deep water shales yielding graptolites, for example at Khekandya River and Lukavy Creek, to biohermal and biogenic–detrital carbonates with mixed brachiopod–coral–graptolite faunas as at Mirny Creek and the Ina and Neznakomka rivers. The upper part of the Tirekhtyakh horizon (the *extraordinarius* and *persculptus* Zones) and the lower part of the Chalmak horizon (the *acuminatus* and *ascensus* Zones) are represented by sequences more

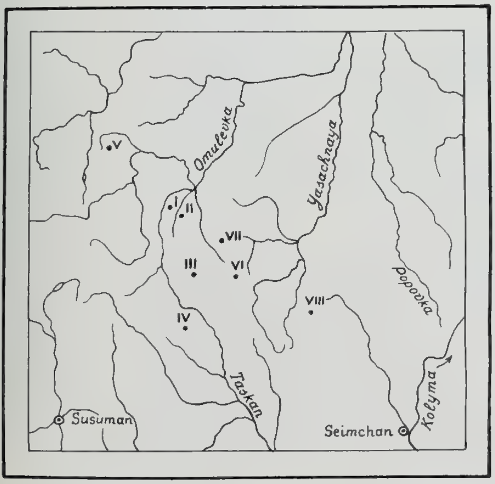

Fig. 1 Distribution of Ordovician–Silurian boundary beds on the Omulev Uplift. I, Mirny Creek; II, Ina River; III, Neznakomka River Basin; IV, Tirekhtyakh River Basin; V, Mount Kharkindzha; VI, Levaya Khekandya River; VII, Drevnyaya River; VIII, Lukavy Creek.

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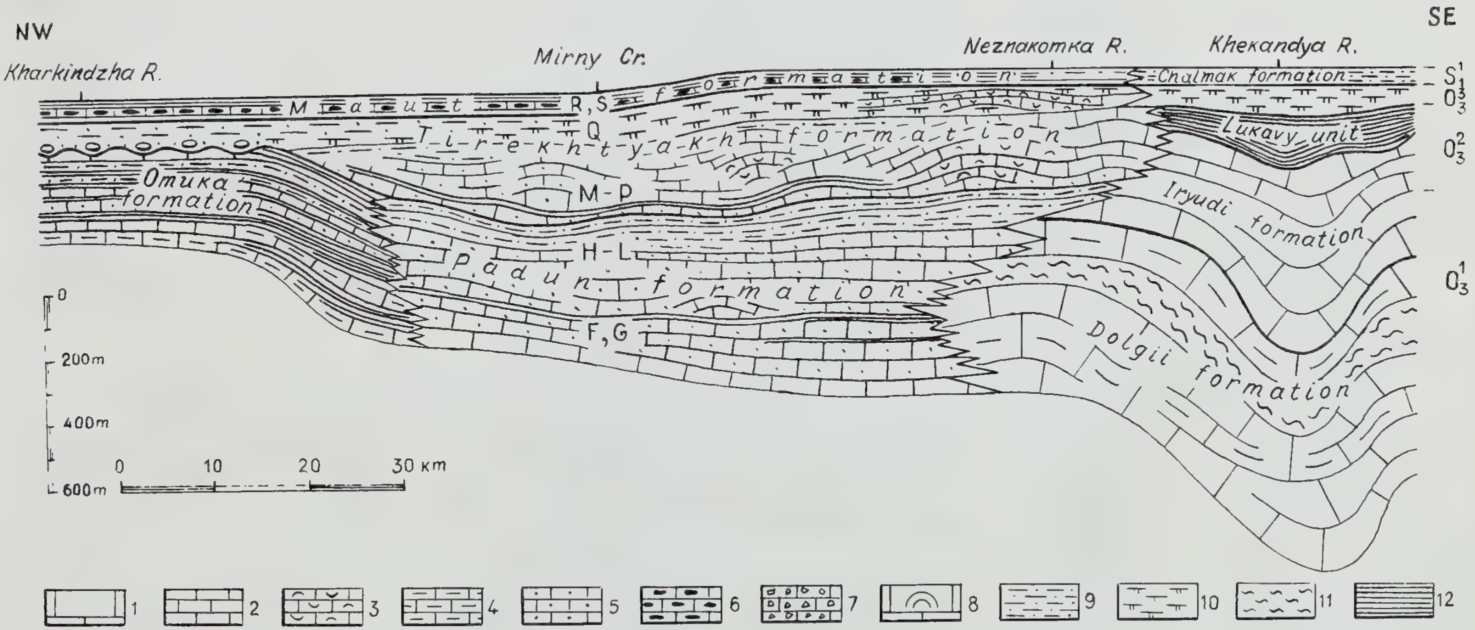
NW
SE
Kharkindzha R.
Mirny Cr.
Neznakomka R.
Khekandya R.
Maut R,S formation
Chalmak formation
Tirekhtyakh formation
Q
Lukavy unit
Omuka formation
M-P
Iryudi formation
H-L
Padun formation
F,G
Dolgii formation
S_1^1
O_3^3
O_3^2
O_3^1
0
200m
400m
600m
0
10
20
30 km
1
2
3
4
5
6
7
8
9
10
11
12

Fig. 2 Lithofacies profile of the Ordovician–Silurian boundary beds on the Omulev Uplift. 1, pelitomorphic massive limestone; 2, pelitomorphic tabular limestone; 3, bioclastic limestone; 4, clayey limestone; 5, silty limestone; 6, siliceous limestone; 7, brecciated limestone; 8, bioherms; 9, calcareous siltstone; 10, dolomitic marl; 11, marl; 12, calcareous shale.

diverse in composition. The Upper Tirekhyakh deposits consist mainly of dolomites, marls, and siltstones representing the termination of the late Ordovician regressive cycle and the Chalmak dark carbonate clay sequences mark the beginning of the Llandovery transgression.

Of greatest interest is the key section at Mirny Creek, which has the best exposed Ordovician–Silurian boundary deposits. This has been studied in detail, and forms a type section for such regional units as formations and horizons.

The Tirekhtyakh horizon

At the Mirny Creek and Ina River sections the horizon is 250 m thick and represented by a formation of the same name (upper unit M to unit Q) which is composed of bedded and massive limestones with tabulate corals, brachiopods, ostracodes and gastropods. The limestones are interbedded with siltstones yielding graptolites. In the Neznakomka River the formation is 315 m thick and represented mainly by biohermal and biogenic-clastic limestones interbedded with siltstones. The rocks contain chiefly brachiopods but the siltstones yield rare graptolites.

In the south-eastern Omulev Mountains (Khekandya River, Yasachnaya Basin, Lukavy Creek and Drevnyaya River) the Tirekhtykh horizon exhibits changes in composition. Its lower part consists of the Iryudi Formation (500–600 m) and Lukavaya sequence (100 m). The Iryudi Formation is composed of clay and pelitomorphic, unevenly bedded limestones with abundant corals and brachiopods. The Lukavaya unit is represented by dark platey limestones intercalated with calcareous shales containing abundant graptolites and rare brachiopods. As at Mirny Creek, the upper part of the horizon includes siltstones.

The Terekhtyakh horizon has been subdivided by means of graptolites in sections at Mirny and Lukavy creeks, the Khekandya and Drevnyaya rivers, and at Mount Kharkindza, and by means of brachiopods mainly in Mirny Creek and the Neznakomka River (Fig. 3). The lower part of the horizon is equated with the *Climacograptus longispinus supernus* Zone and the *Tcherskidium unicum* beds. The *supernus* Zone is subdivided into two subzones, the lower *Climacograptus longispinus longispinus* Subzone and the upper *Paraorthograptus pacificus* Subzone. The lower subzone contains *Climacograptus longispinus longispinus* Hall, *C. l. supernus* Elles & Wood, *C. hastatus* Hall, *C. trifidus spectabilis* Koren & Sobolevskaya, and *Dicellograptus complanatus* Lapworth, whose appearance marks its lower boundary. The *pacificus* Subzone is recognized as a taxon biozone and, along with *Dicellograptus ornatus ornatus* Elles & Wood and subspecies of *Climacograptus longispinus*, contains *Climacograptus latus hekandaensis* Koren & Sobolevskaya and *C. pogrebovi* Koren & Sobolevskaya, while the upper part yields *Glyptograptus*? *ojsuensis* Koren & Mikhailova, *Climacograptus angustus* (Perner), *C. normalis* Lapworth and others.

The *supernus* Zone is equated with the *Tcherskidium unicum* beds which also contain *Ptychoglyptus bellarugosus* Cooper, *Holorhynchus* ex gr. *giganteus* Kiaer and *Eostropheodonta hirnantensis lucavica* Oradovskaya. There are also abundant corals of the genera *Agetolites*, *Heliolites*, *Propora*, *Calapoecia*, *Coxia* and others (Preobrazhensky 1966). The brachiopod-coral assemblage allows the lower Tirekhtyakh horizon to be correlated with the 5b beds of Norway. On Mirny Creek the deposits also contain trilobites, gastropods, ostracodes and other fossils (Sokolov *et al.* 1983).

The upper Tirekhtyakh horizon corresponds to the *Climacograptus*? *extraordinarius* and *Glyptograptus*? *persculptus* Zones. The *extraordinarius* Zone, which was first established on Mirny Creek (Koren & Sobolevskaya 1979), corresponds to the index-species range. Apart from the latter, it contains *Climacograptus*? ex gr. *extraordinarius* (Sobolevskaya), *C. angustus* (Perner), *C. normalis* Lapworth and *C. mirnyensis* (Obut & Sobolevskaya). *Climacograptus* aff. *medius* Törnquist and scarce *Glyptograptus* sp. appear in the upper part of the zone.

The *persculptus* Zone was recognized as equal to the full range of the index-species and the zonal assemblage also contains *Climacograptus angustus* (Perner), *C. normalis* Lapworth, *C. mirnyensis* (Obut & Sobolevskaya) and *C. torosus* Koren & Sobolevskaya.

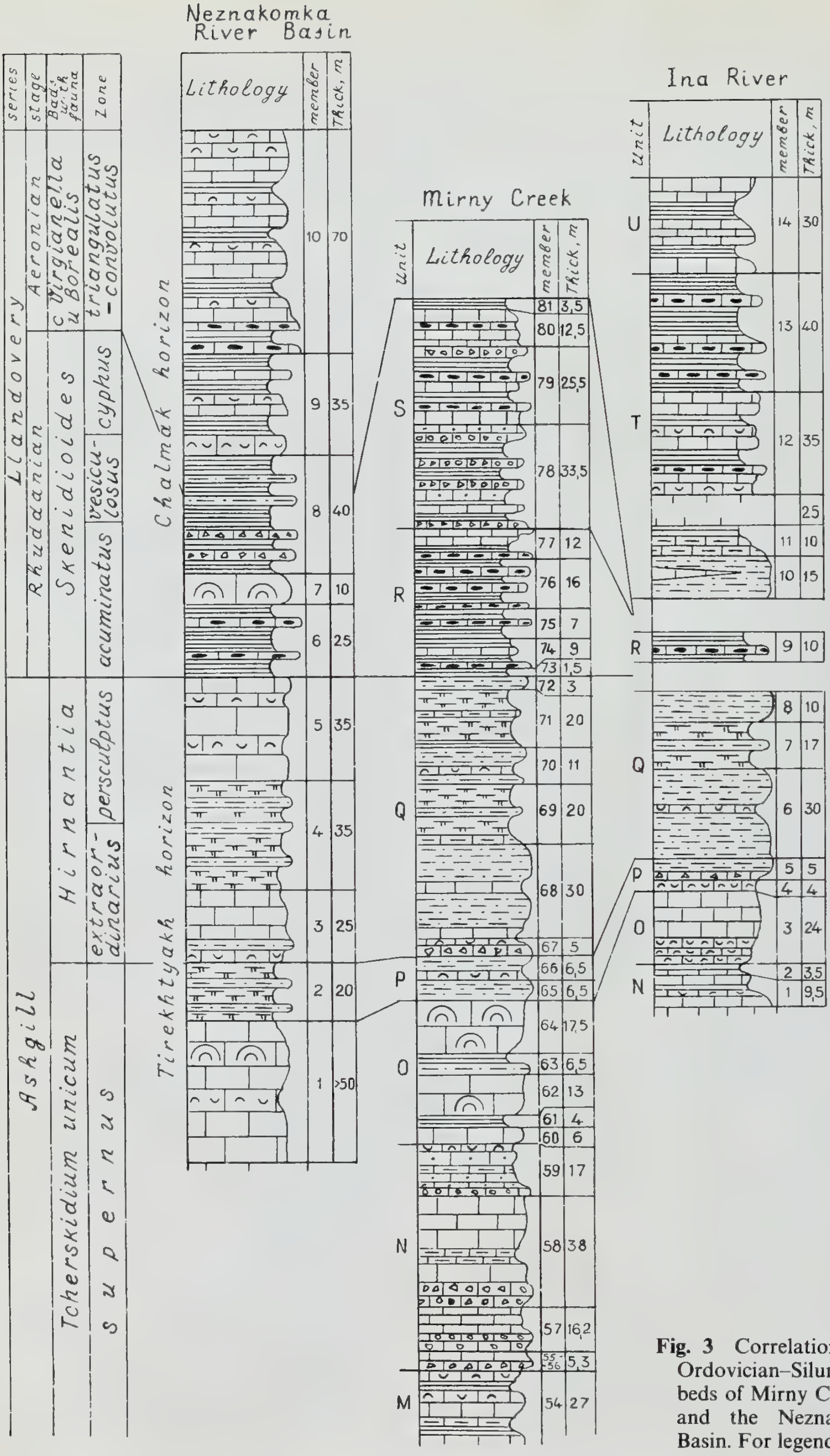
Neznakomka River Basin
series
stage
Beds with fauna
Zone
Lithology
member
Thick, m
Llandovery
Aeronian
Rhuddanian
Askgill
c Virgianella u Borealis
Skenidioides
Hirnantia
Tcherskidium unicum
triangulatus -convolutus
cyphus
vesiculosus
acuminatus
persculptus
extraordinarius
supernus
Chalmak horizon
Tirekhtyakh horizon
10 70
9 35
8 40
7 10
6 25
5 35
4 35
3 25
2 20
1 >50
Mirny Creek
unit
Lithology
member
Thick, m
S
R
Q
P
O
N
M
81 3,5
80 12,5
79 25,5
78 33,5
77 12
76 16
75 7
74 9
73 1,5
72 3
71 20
70 11
69 20
68 30
67 5
66 6,5
65 6,5
64 17,5
63 6,5
62 13
61 4
60 6
59 17
58 38
57 16,2
55-56 5,3
54 27
Ina River
unit
Lithology
member
Thick, m
U
T
R
Q
P
O
N
14 30
13 40
12 35
25
11 10
10 15
9 10
8 10
7 17
6 30
5 5
4 4
3 24
2 3,5
1 9,5

Fig. 3 Correlation chart of the Ordovician–Silurian boundary beds of Mirny Creek, Ina River and the Neznakomka River Basin. For legend see Fig. 2.

At Mirny Creek, the Khekanda and Neznakomka rivers and Mount Kharkindzha, this zone is equated with the *Hirnantia*? beds (Oradovskaya 1977). Amongst the brachiopods the most common are *Dolerorthis*? *savagei* Amsden, *Brevilamunella thebesensis* (Savage), *Rafinesquina*? *latisculptilis* (Dalman) and *Giraldibella bella* (Bergström), and the trilobites *Bumastus* (*Bumastus*) *commodus* Apollonov and *Mucronaspis kolymica* Chugaeva. *Dalmanitina olini* Temple occurs near the top of the zone.

The Chalmak horizon

In the Omulev Mountains the lower Chalmak horizon includes the Maut Formation, and the main Chalmak Formation corresponds to the horizon in the Yasachnaya Basin. On the Omulev Uplift, the Maut Formation consists of dark calcareous shales, shales and cherts containing graptolites which are interbedded with detrital and conglomerate-like limestones with a scarce neritic fauna. Coarse clastic rocks dominate the coeval deposits further south-east.

The lower part of the horizon corresponds to the *Parakidograptus acuminatus* and *Akidograptus ascensus* Zones recognized in Mirny Creek, the Ina and Khekanda rivers, and Mount Kharkindzha. The most complete graptolite assemblage was reported from Mirny Creek (Obut *et al.* 1967). As well as *P. acuminatus* and *A. ascensus*, the assemblage includes *Climacograptus rectangularis* (M'Coy), *C. transgrediens* Waern, *Paraclimacograptus sinitzini* Chalatskaya, *Diplograptus* ex gr. *modestus* Lapworth and *Glyptograptus* ex gr. *tamariscus* (Nicholson). The boundary of the zone is drawn by the appearance and disappearance of the diagnostic species.

The *acuminatus* and *ascensus* Zone corresponds to the lower *Skenidioides* beds containing *Skenidioides* cf. *scolioides* Temple, *Leptaena* aff. *aequalis* Amsden, *Eospirigerina putilla* Oradovskaya, *Zygospiraella* sp. and *Protatrypa* sp. The assemblage is similar to the brachiopod fauna from the lower Llandovery of the Northern Appalachians (Ayrton *et al.* 1969). The beds also contain trilobites such as *Acernaspis* sp., Tropidocoryphinae gen. et sp. indet. and the corals *Palaeofavosites balticus* Rukhin, and *Propora conferta* Edwards & Haime, among others.

The systemic boundary

The most complete and well known section of the Tirekhtyakh and Chalmak horizons is exposed along the Mirny Creek. A point 2·5 km from its mouth was chosen as a regional type section for the Ordovician–Silurian boundary in the north-east USSR. The systemic boundary is drawn at the base of unit 73 which is 1·5 m thick and coincides with the base of the Maut Formation (Figs 2, 3). This level corresponds to the base of the *acuminatus* and *ascensus* Zone which in the section studied is substantiated by the appearance of representatives of such typically Silurian groups as *Diplograptus modestus* Lapworth and *Glyptograptus tamariscus* (Nicholson) (unit 73). The index-species *Akidograptus ascensus* Davies is known from the base of unit 74, 1·5 m above the boundary, and *Parakidograptus acuminatus* (Nicholson) occurs in the lower part of unit 75, 11 m above the boundary. Their absence from the basal layer can be attributed to the difficulty in searching for graptolites in the beds. In the section at Mount Kharkindzha, akidograptids are known from the basal beds of the Maut Formation associated with other typical diplograptids.

The principal criteria for establishing the boundary on a regional scale are distinct changes in the lithological composition of the deposits as well as the change in the assemblages of graptolites (*persculptus*/*acuminatus* and *ascensus*), brachiopods (*Hirnantia*?/*Skenidioides*) and trilobites. Graptolites allow interregional and global correlations of the level.

The section at Mirny Creek is well exposed and shows a continuous succession of uniformly dipping deposits containing diverse fossils. Its major advantage is bed-by-bed graptolite control within the range of the *Dalmanitina–Hirnantia* assemblage and the presence of shelly fauna (the *Skenidiodes* beds) from the base of the *acuminatus* and *ascensus* Zones.

Abundant graptolites, brachiopods and corals and rare trilobites, ostracodes and conodonts are known from the Ordovician–Silurian boundary beds in the north-east USSR. All faunal groups except ostracodes and conodonts have been monographically described in different publications (Sokolov *et al.*, 1983; Nikolaev *et al.*, 1977; Nikolaev & Sapelnikov 1969; Obut *et al.* 1967; Opornii razrez (Anon.) 1974; Oradovskaya 1963; Polevoi atlas (Anon.) 1968; Polevoi atlas (Anon.) 1975; Preobrazhensky 1966 and Sobolevskaya 1970, 1974).

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The Ordovician–Silurian boundary in the Altai Mountains, USSR

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Synopsis

The Ordovician–Silurian boundary is repeatedly exposed in the Altai–Sayan fold belt, with the best-studied outcrops in the Charysh–Inya structural zone near Ust'-Chagyrka and Chineta, where the *persculptus* and *acuminatus* zones are both known in association with shelly faunas.

Ordovician and Silurian deposits in the western part of the Altai–Sayan fold-belt are not only widely distributed in the Altai, but in the Kuznetsk Alatau, Salair and Shoria Mountains as well. The boundary interval, however, is known only from the Altai Mountains in two structural-formational zones, the Anui–Chuya and Charysh–Inya zones. In the first zone there are several sections where it is possible to see a normal stratigraphical succession from Ordovician to Silurian. However, most of them are not well characterized palaeontologically, especially the boundary beds (Yolkin *et al.* 1978; Sennikov & Sennikov 1982). Because of this, the boundary interval is shown as a biostratigraphical break in the stratigraphical correlation charts for this zone (Khomentovskiy & Tesakov 1983).

The Ordovician–Silurian boundary interval is better known in the Charysh–Inya Zone. Here, in different areas, there are now more than ten known sections. In each such area there are usually several sections with transitional continuity between the two systems, though there are some differences in the faunas from area to area. The best of these sections occur near Ust'-Chagyrka and Chineta villages (Yolkin & Zheltonogova 1974; Sennikov *et al.* 1979, 1982, 1984). The faunal assemblages in these sections in the two areas include graptolites, conodonts, trilobites, gastropods, orthoconic cephalopods, brachiopods, ostracodes, corals, chitinozoans and polychaetes, part of which have been monographed (Sennikov 1976, 1978; Moskalenko 1977; Severgina 1978, 1984; Yolkin 1983). The most important fossils for the subdivision and correlation of these sections are the graptolites. They are the predominant group numerically and have by far the best international distribution stratigraphically.

It is important to draw attention to the association, in the boundary beds, of graptolites, conodonts and trilobites, especially *Dalmanitina*. This indicates the possibility for future work directed towards clarifying and refining the correlation of Ordovician–Silurian boundary beds in the Altai, but perhaps also globally. The best boundary in the Altai, as in China (Chen Xu 1984) would be somewhat below the *acuminatus* Zone decided by the Ordovician–Silurian Boundary Working Group (Holland 1985). The beds with *persculptus* correspond to the onset of a wide transgression.

Bull. Br. Mus. nat. Hist. (Geol) **43**: 139–143

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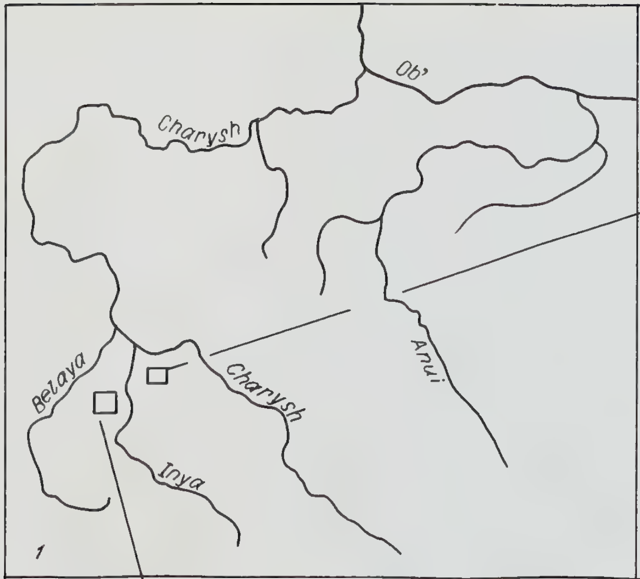
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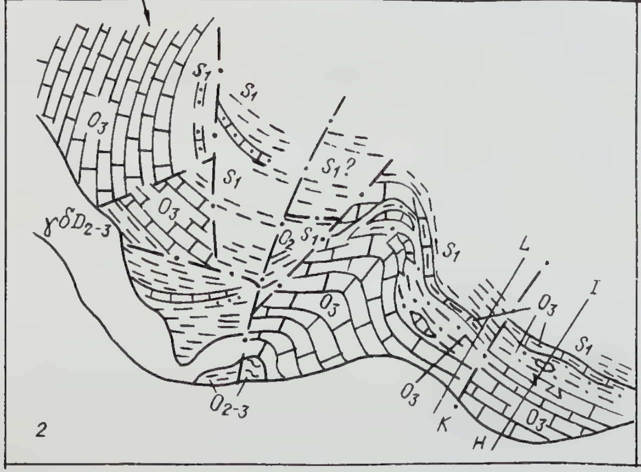
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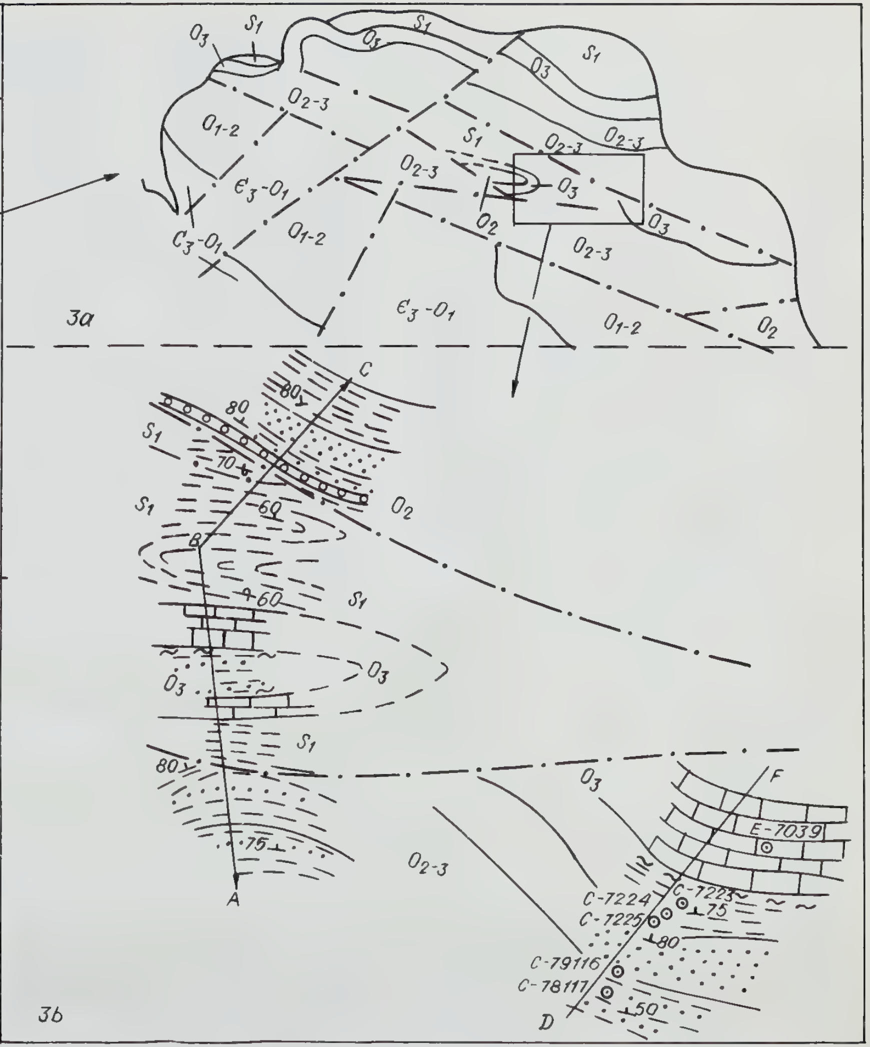
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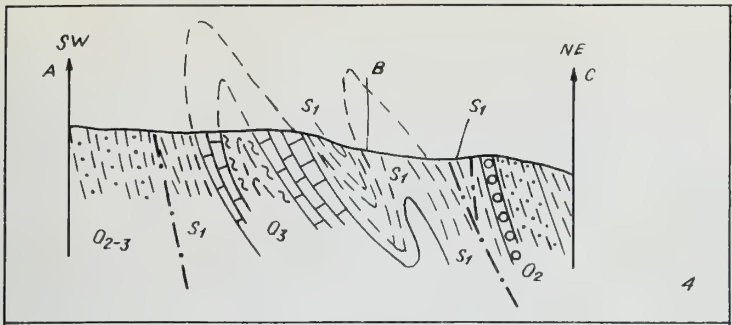
E-7039
C-7223
C-7224
C-7225
C-79116
C-78117

Fig. 1 Location of reference sections of Ordovician and Silurian boundary beds in the Altai Mountains.
Fig. 2 Geological map of the left bank of Buroblyanka Creek near Chineta village.
Fig. 3 Geological map of the left bank of Chagyrka Creek : 3a—sketch map, 3b—details of section areas.


SW
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O_3
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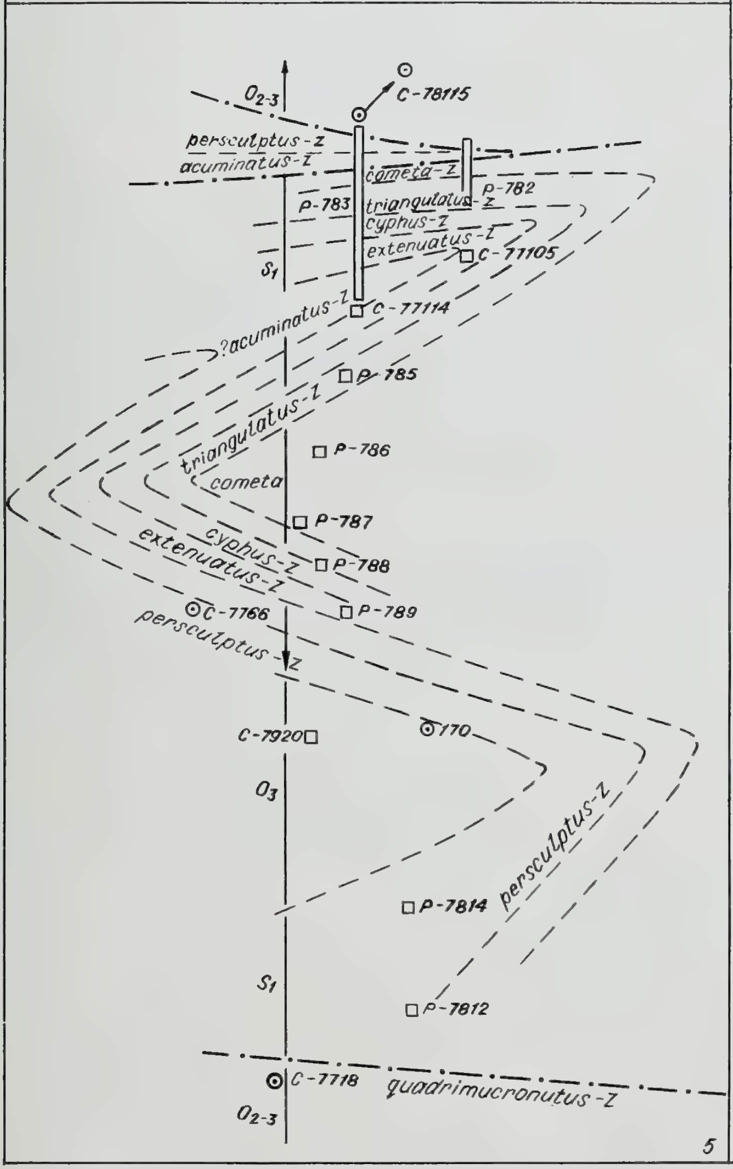
4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C-77105
S_1
C-77114
?acuminatus-Z
P-785
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cometa
P-787
cyphus-Z
P-788
extenuatus-Z
C-7766
P-789
persculptus-Z
C-7920
770
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O_{2-3}
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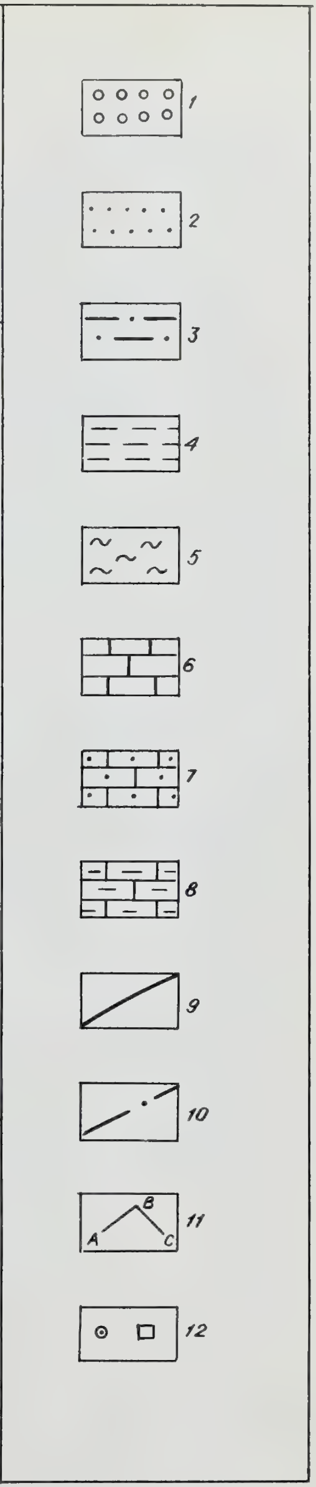
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Fig. 4 Profile on line A–B–C of Fig. 3b.

Fig. 5 Details of graptolite zonation on line A–B–C of Fig. 3b.

Legend for Figs 2–7. 1—conglomerates, 2—sandstones, 3—silty sandstones, 4—siltstones, 5—cherty rocks, 6—limestones, 7—sandy limestones, 8—argillaceous limestones, 9—boundaries, 10—faults, 11—line of sections, 12—outcrops and artificial exposures (fauna collection points).

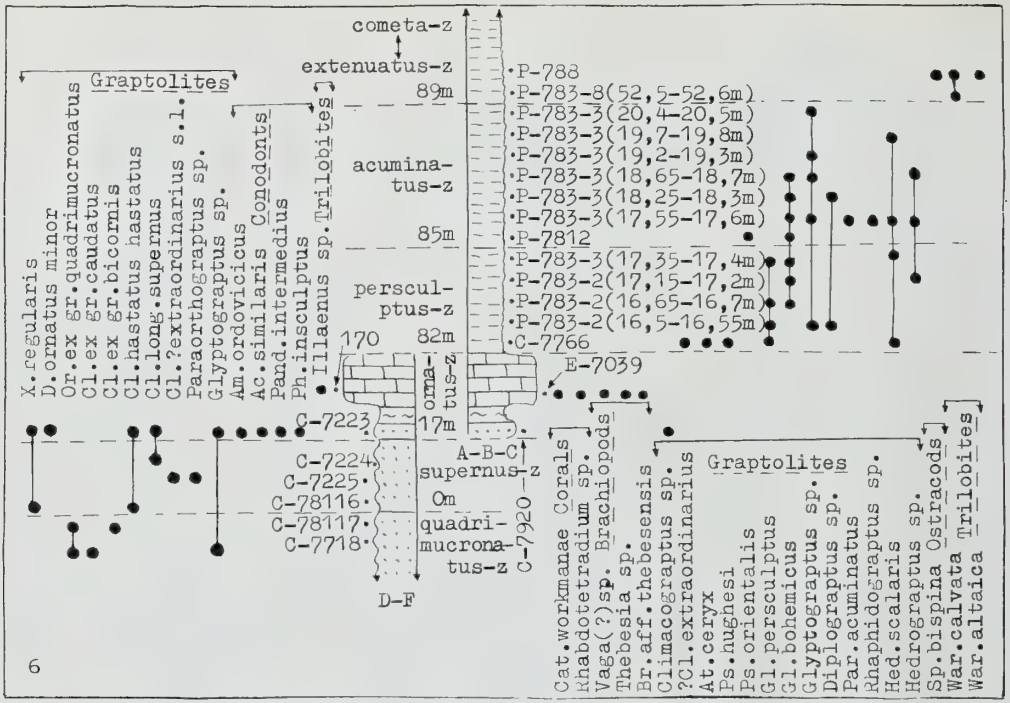

Fig. 6 Distribution of faunas in sections near Ust'-Chagyrka village.

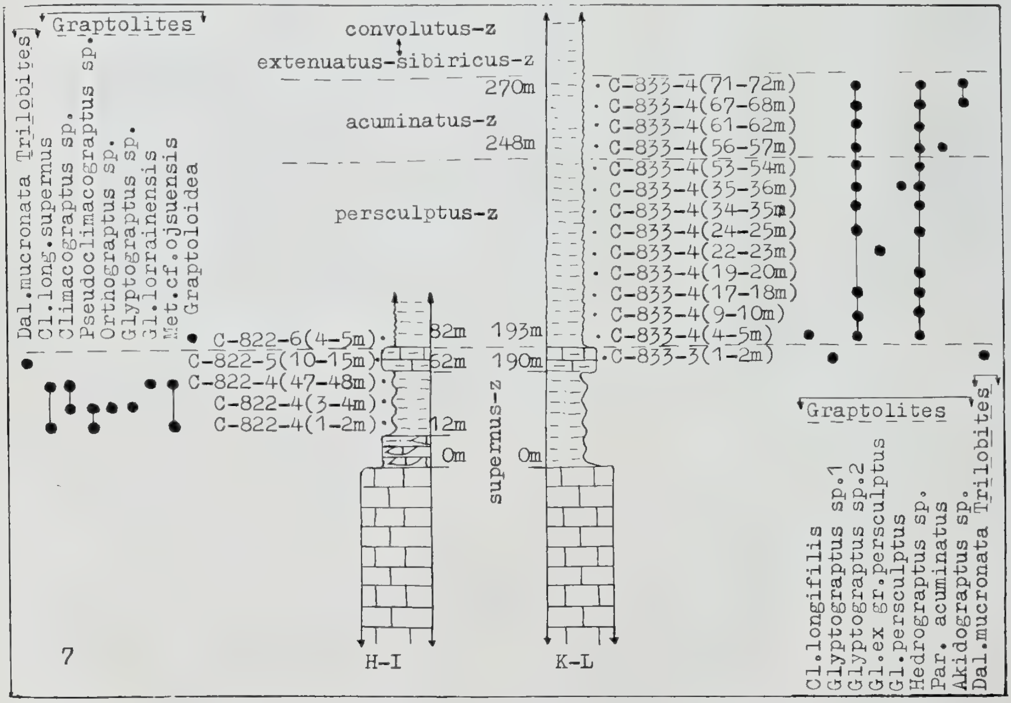

Fig. 7 Distribution of faunas in sections near Chineta village.

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Nature of the Ordovician–Silurian boundary in south Kazakhstan, USSR

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Synopsis

Kazakhstan was the region where the coeval nature of the *Dalmanitina mucronata–Hirnantia* faunas with the *persculptus* Zone faunas was first established. The best sections are in the Chu-Ili Mountains of South Kazakhstan, the Ashchisu River and the Zhideli and Karasay sequences. A summary is given of the upper Ashgill and lower Llandovery biostratigraphy and the position of the systemic boundary. The lithostratigraphy is also outlined.

To have the Ordovician–Silurian boundary at the base of the *acuminatus* Zone was first advanced by Kazakhstan geologists (Rukavishnikova *et al.* 1968; Mikhailova 1970; Nikitin 1972; Apollonov *et al.* 1973; Apollonov 1974; Poltavtseva & Rukavishnikova 1972) after the discovery of *Glyptograptus persculptus* in association with *Dalmanitina mucronata* and *Hirnantia* in the Chu-Ili Mountains. This showed that the *persculptus* Zone did not succeed the *Dalmanitina* beds, as was previously thought in western Europe, and that, on the contrary, it was partly coeval with the *Dalmanitina mucronata–Hirnantia* beds which have always been assigned to the Ordovician. Thus it became clear that tracing the *persculptus* boundary in the neritic facies was impossible. This new evidence has been widely discussed in the literature (Williams *et al.* 1972; Bergström *et al.* 1973; Lespérance 1974; Rozman 1976; Rickards 1976).

The Kazakhstan Ordovician–Silurian boundary deposits are best studied in the Chu-Ili Mountains in south Kazakhstan, in the upper reaches of the Ashchisu River (Durben and Ojsu wells), as well as along the Zhideli and Karasay dry channels (Apollonov *et al.* 1980; Nikitin *et al.* 1980: textfigs 1–6). This paper is a summary of the upper Ashgill and lower Llandovery biostratigraphy and describes the position of the system boundary established in Kazakhstan on the basis of continuous sections.

The succession is divided into three conformable lithostratigraphic units: the Chokpar, Zhalair and Salamat Formations. The latter is overlain by the Betkainar Formation (Figs 1–6).

The Chokpar Formation consists of dark-grey and greenish-grey regularly bedded mudstones and siltstones yielding abundant graptolites characteristic of the *supernus* Zone (Apollonov *et al.* 1980). A more detailed zonation can now be suggested. The lowermost part of the Chokpar Formation contains *Dicellograptus ornatus minor* Toghill, *Climacograptus longispinus supernus* Elles & Wood, *Amplexograptus inuiti* (Cox) and *Orthograptus amplexicaulis* (Hall) and comprises the *inuiti* Zone. The graptolites present above this, and in most of the Chokpar Formation, are characteristic of the *pacificus* Zone and include *Dicellograptus ornatus* Elles & Wood, *Climacograptus manitoulinensis* Caley, *Orthograptus socialis* (Lapworth), *Paraorthograptus pacificus* (Ruedemann) (rare) and *Nymphograptus velatus* Elles & Wood. The uppermost Chokpar Formation locally contains limestone beds which are best developed in the Osju section where they are placed in a local stratigraphic unit—the Osju Limestones. The unit consists of dark-grey argillaceous limestones interbedded with aphanitic sandy limestones in which terrigenous clastics account for 15 to 20%. The Osju Limestones yields abundant brachiopods and trilobites including *Giraldiella bella* Bergström, *Streptis altosinuata* (Holtedahl), *Leptaena rugosa* Dalman, *Cryptothyrella* sp., *Tscherskidium* cf. *ulkuntasensis* Sapelnikov & Rukavishnikova, *Prostricklandia prisca* Rukavishnikova & Sapelnikov, *Platycoryphe sinensis sinensis* (Lu),

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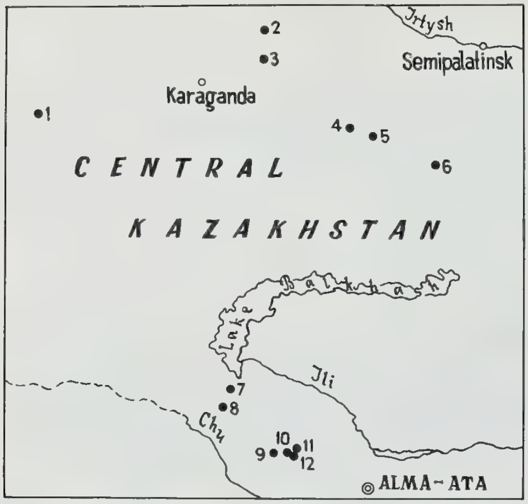


Fig. 1 Localities of the Ordovician–Silurian boundary deposits in Central and South Kazakhstan. 1, Sarysu–Teniz watershed and Zhaksykon River; 2, Northeast of Central Kazakhstan–Kombabasor lake; 3, Akjar–Zhartas watershed; 4–6, Chingiz Range and Pre-Chingiz Range: 4, Mount Otyzbes; 5, Mount Mizek; 6, Mount Akdombak; 7–12, Chu-Ili Mountains: 7, Karasay River; 8, Zhideli River; 9, Anzhar River; 10, Ojsu well; 11, Durben well; 12, Mount Dulankara.

Bumastus commodus Apollonov, *Decoroproetus artus* Apollonov, *D.* cf. *evexus* Owens, *Otarion curvulum* Apollonov, *O. gibberum* Apollonov, *Dicranogmus confinis* Apollonov, and *Leonaspis* sp. There also occur conodonts, bivalves, gastropods and cepalopods, among them *Acodus similaris* Rhodes, *Eobelodina fornicala* Stauffer, *Icriodella* sp., *Tshuiliceras lobatum* Barskov, *Michelinoceras procurens* Barskov and *Geisonoceras fustis* Barskov.

The numerous graptolites that are characteristic of the *pacificus* Zone occur in argilliceous limestone layers. Present are *Climacograptus longispinus supernus* Elles & Wood, *C.* cf. *normalis* Lapworth, *C. tatianae* Keller, *Glyptograptus posterus* Koren & Tzai, *G.*? *ojsuensis* Koren & Mikhailova, *Paraorthograptus pacificus* (Ruedemann), *Orthograptus amplexicaulis* (Hall) and *Plegmatograptus nebula lautus* Koren & Tzai. Rare tabulate corals, radiolarians and algae are also known (Apollonov *et al.* 1980).

The uppermost Chokpar Formation in other sections (as at the Anzhar River) is represented by massive biogenic-detrital limestones (the so-called Ulkuntas Limestones) overcrowded with tabulate corals and heliolitids. The assemblage includes *Agetolites* cf. *mirabilis* Sokolov, *Hemiagetolites insignis* Poltavceva, *Catenipora inordinata* Kovalevsky, *Plasmoporella papillatiformis* Kovalevsky, *Propora cancellatiformis* Sokolov and *Heliolites parvulus* Kovalevsky. Some pentamerids such as *Holorhynchus giganteus* Kiaer, *Proconchidium tchuilensis* Rukavishnikova & Sapelnikov and *Tcherskidium*? *ulkuntasense* Rukavishnikova & Sapelnikov have been found. There occur the trilobites *Holotrachellus punctillosus* Törnquist, *Amphylicas* sp. and *Sphaerexochus* sp., which are characteristic of biohermal environments. The thickness of the Ojsu and Ulkuntas Limestones varies from 14 to 55 m and the Chokpar Formation totals 140 to 180 m.

The Zhalair Formation rests conformably on the Chokpar deposits and is exposed in all sections studied. The section at Durben may serve as a stratotype (Figs 2, 3). The formation is composed of tobacco-green and greenish-grey siltstones interbedded locally with grey and reddish-brown fine-grained poorly sorted sandstones, the latter being of carbonate and quartz-feldspathic composition. Locally, sandstones form a separate unit more than 80 m thick, for example at the Ojsu section. The lowermost Zhalair Formation includes the Durben Limestone which is 9 to 40 m thick, and is easily discernible in many of the sections studied (Fig. 4). It consists of well-bedded dark grey pelitomorphic limestones. The upper part of the Zhalair Formation contains local beds of dark grey and green silty tuffites.

The lower Zhalair Formation (the Durban horizon) contains graptolites of the *extraordinarius* and *persculptus* Zones (Koren & Nikitin 1983). The former zone yields *Climacograptus angustus* (Perner), *C. normalis* Lapworth, *C.*? *extraordinarius* (Sobolevskaya) (= *Glyptograptus*? *persculptus* forma A and *G.* aff. *persculptus* of Apollonov *et al.* 1980) and *Pseudoclimacograptus*

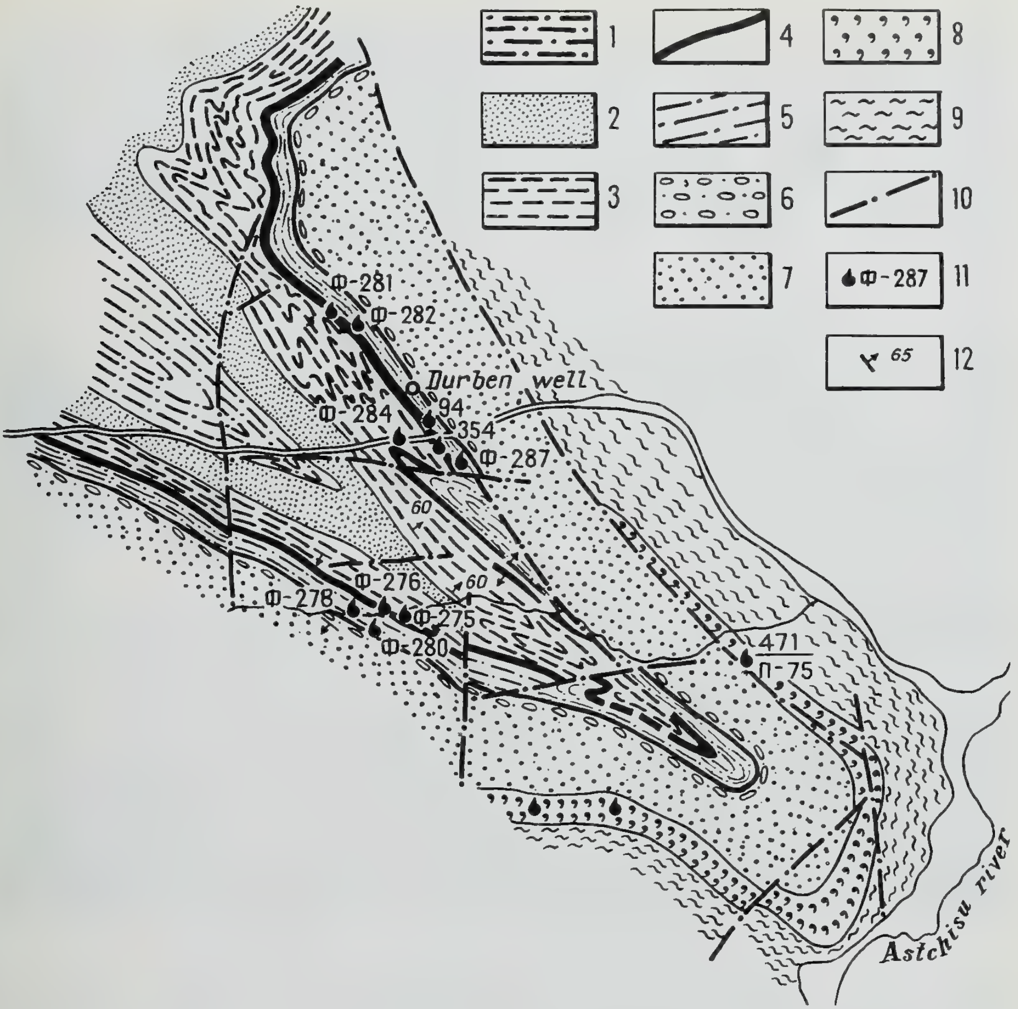


Fig. 2 Schematic geological map of the Durben well area. 1, 2, Kysylsai Formation (?): 1, black siltstones and sandstones; 2, yellow sandstones; 3, Chokpar Formation black mudstones and siltstones; 4, 5, Zhalair Formation: 4, dark fine-crystalline and fine-clastic limestones; 5, green siltstones and fine-grained sandstones; 6, 7, Betkainar Formation: 6, basal conglomerate and sandstones; 7, grey sandstones; 8, red sandstones; 9, Koichin Formation: red sandstones and siltstones; 10, faults; 11, localities of fauna; 12, strike and dip.

sp. The latter zone may be distinguished by the occurences of *Glyptograptus persculptus* (Salter) (=*G. persculptus* forma B of Apollonov *et al.* 1980), *Glyptograptus* sp. and *Climacograptus angustus* (Perner). A shelly fauna was found in limestones and siltstones within both graptolite zones, namely a typical *Dalmanitina–Hirnantia* assemblage including *Platycoryphe sinensis* (Lu), *Dalmanitina mucronata* (Brongniart), *Dalmanitina olini* Temple, *Leonaspis olini* Troedsson, *Dicranopeltis* sp., *Dalmanella testudinaria* (Dalman), *Hirnantia sagittifera* (M'Coy), *Anisopleurella novemcostata* Nikitin, *Aegiromena durbenensis* Nikitin, *Aphanomena ultrix* (Marek & Havlíček), *A.* aff. *urbicola* (Marek & Havlíček), *Bracteoleptaena polonica* Temple, *Eostropheodonta bublitschenki* Nikitin and *Coolinia iliensis* Nikitin.

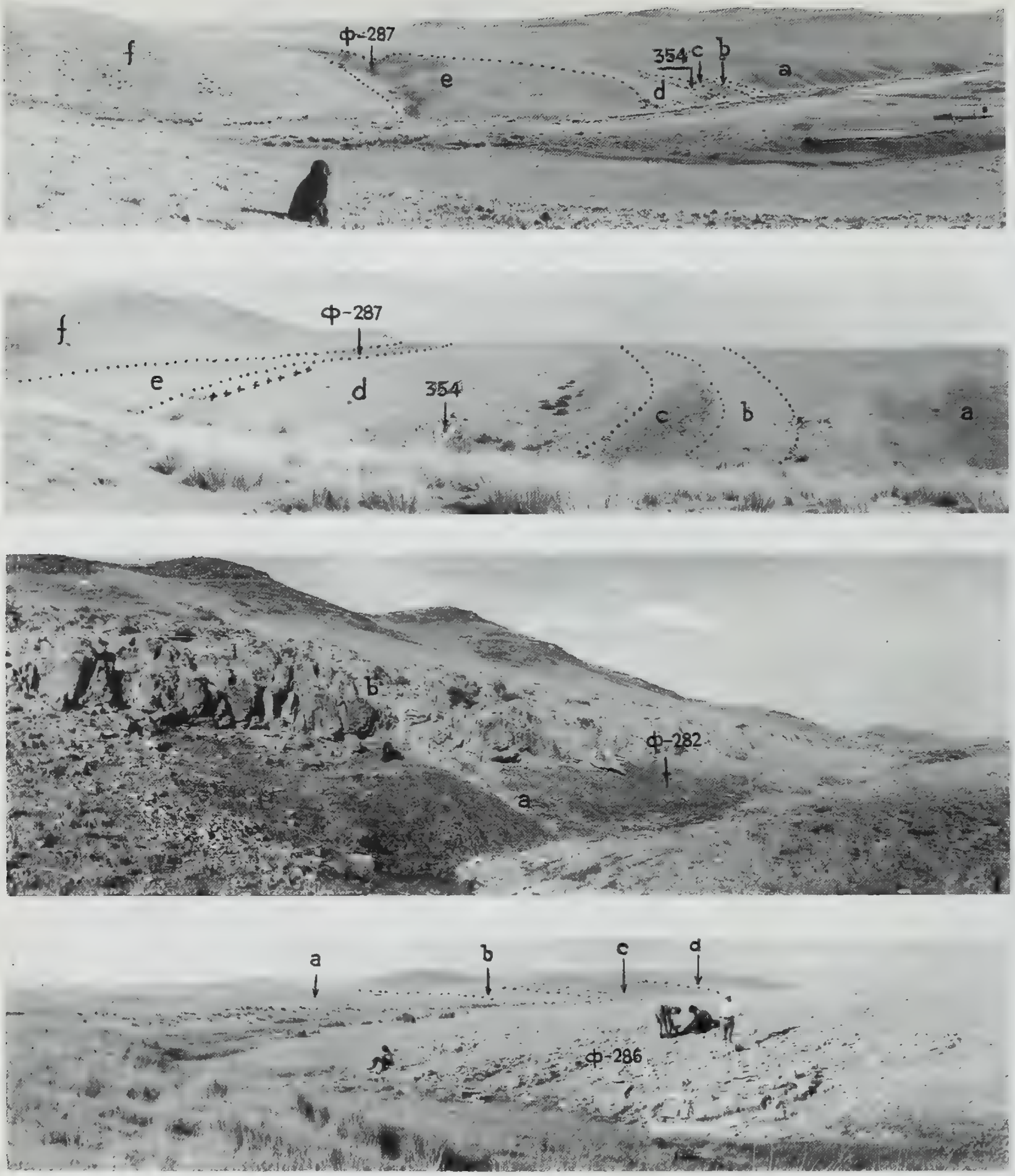


Fig. 3 A—Section on the north-east limb of the Ashchysu anticline near the Durben well. (a) the Chokpar Formation, (b–d) the Zhalair Formation: (b) limestones, (c) carbonaceous sandstone, (d) limestones, (e) siltstones, (f) Betkainar Formation; 354, f-287—localities of fauna. In the background to the right are hills composed of coarse-grained sandstones of Betkainar Formation on the south-western limb of the anticline.
B—enlarged part of the same section.
C—section near the Ojsu well. (a) Ojsu Limestones of the uppermost part of the Chokpar Formation; (b) limestones with *Dalmanitina* assemblage; (c) siltstone of the basal Silurian. In the foreground an exposure of the Ojsu Limestones is seen.
D—transgressive onlapping of the basal conglomerate of the Betkainar Formation (b) on siltstones of the middle Zhalair Formation (a) in the Durben well area. Photographs I. F. Nikitin.

The thickness of the lower Zhalair Formation (the *extraordinarius* and *persculptus* Zones) varies from 122 to 127 m in the southeastern Chu-Ili Mountains (the Durben and Osju wells), to 55 m in the Zhideli River and to half a metre in the Karasay River in the northwestern Chu-Ili Mountains.

The upper Zhalair Formation (the Alpeis horizon) yields early Silurian graptolites. The *acuminatus* Zone is well defined in the strata overlying the *persculptus* Zone in sections in the Karasay, Zhideli, and Ashchysu Rivers. The zonal assemblage includes abundant graptolites, namely *Climacograptus acceptus* Koren & Mikhailova, *C.? jidelensis* Koren & Mikhailova, *C. mirnyensis* (Obut & Sobolevskaya), *C.* ex gr. *normalis* Lapworth, *Pseudoclimacograptus* (*Metaclimacograptus*) *fidus* Koren & Mikhailova, *P.* (*M.*) *pictus* Koren & Mikhailova, *Diplograptus modestus primus* Mikhailova, *G. madernii* Koren & Tzai, *Akidograptus* cf. *ascensus* Davies, *A. ascensus cultus* Mikhailova, *Parakidograptus* cf. *acuminatus* (Nicholson) and *Orthograptus illustris* Koren & Mikhailova.

The younger beds of the Zhalair Formation are eroded over most of the area studied (Fig. 4) and they are exposed only in the lower Karasay River. There, in beds overlying the *acuminatus* Zone, the graptolites *Climacograptus miserabilis* Elles & Wood, *Glyptograptus* sp. and abundant *Priblylograptus* sp. and *Atavograptus* sp., characteristic of the *vesiculosus* Zone, were found. The section is capped by strata yielding *Climacograptus angustus* (Perner), *C. mirnyensis* (Obut & Sobolevskaya), *C. normalis* Lapworth, *Pseudoclimacograptus* (*Metaclimacograptus*) *hughesi* (Nicholson), *Coronograptus cyphus* (Lapworth), *C. gregarius* (Lapworth), *Monograptus revolutus praecursor* Elles & Wood, *Atavograptus* sp. and *Dimorphograptus dessicatus* Elles & Wood. Shelly fauna is scarce in the Silurian part of the Zhalair Formation. In the *acuminatus* Zone only a single trilobite of the family Odontopleuridae occurs (exposure 280). The Zhalair Formation is 51 to 133 m thick.

The Salamat Formation consists of green sandstones and siltstones with abundant graptolites of the *gregarius* Zone. The overlying Betkainar Formation, with basal conglomerate beds, transgresses deposits of different ages, including in places the *Dalmanitina mucronata* beds of the Durben horizon (Figs 2, 4).

The Ordovician–Silurian boundary in the Chu-Ili Mountains is drawn at the base of the *acuminatus* Zone, which is marked by the appearance of *Akidograptus ascensus* Davies, *Glyptograptus madernii* Koren & Tzai, *Orthograptus illustris* Koren & Mikhailova and *Diplograptus modestus primus* Mikhailova.

The Chokpar and Zhalair Formations reflect a distinct regressive-transgressive cycle (Fig. 5). Dark pelitomorphic deposits of the Chokpar Formation (the *supernus* Zone) are comparatively deep-water and might have accumulated in an extensive, open, flat-bottomed sea with a remote source of terrigenous sediments. That sea was inhabited by diverse graptolites (more than 15 species). Towards the end of Chokpar time, the sea bed was elevated and a number of biohermal chains were developed. Each bioherm had a trail of clastic carbonate material (the Ulkuntas Limestones).

In early Durben time (the *extraordinarius* Zone), the areas of continuously growing elevation were surrounded by thick beds chiefly consisting of limey coarse-grained sands (Fig. 6), and a broad band of the fine dark Durben Limestones accumulated which were 40 m thick near the elevations and 0·5 m thick further away. The areas of limestone sedimentation were inhabited by a trilobite assemblage including *Dalmanitina mucronata*, *D. olini* and *Platycoryphe sinensis.* In the deep-water limestones near the village of Karasay a single species of blind *Dalmanitina* was found. Brachiopods are commonly represented by the single species *Bracteoleptaena polonica.* The graptolite assemblage consists of 2 to 4 species, and all the fossils are large-sized, numerous but taxonomically restricted. Late Durben time (the *persculptus* Zone) saw the deposition of green fine-grained sandstones and cross-bedded siltstones with traces of turbidity and slumping. The benthic fauna shows a greater diversity (the *Hirnantia–Dalmanitina* assemblage) but the graptolites are limited to two to three species.

An abrupt increase in the supply of tuffaceous material coincided with the beginning of the *acuminatus* Zone. A new and diverse (up to 15 species) graptolite assemblage appeared; however, benthic faunas are almost unknown from this level. The cosmopolitan distribution of

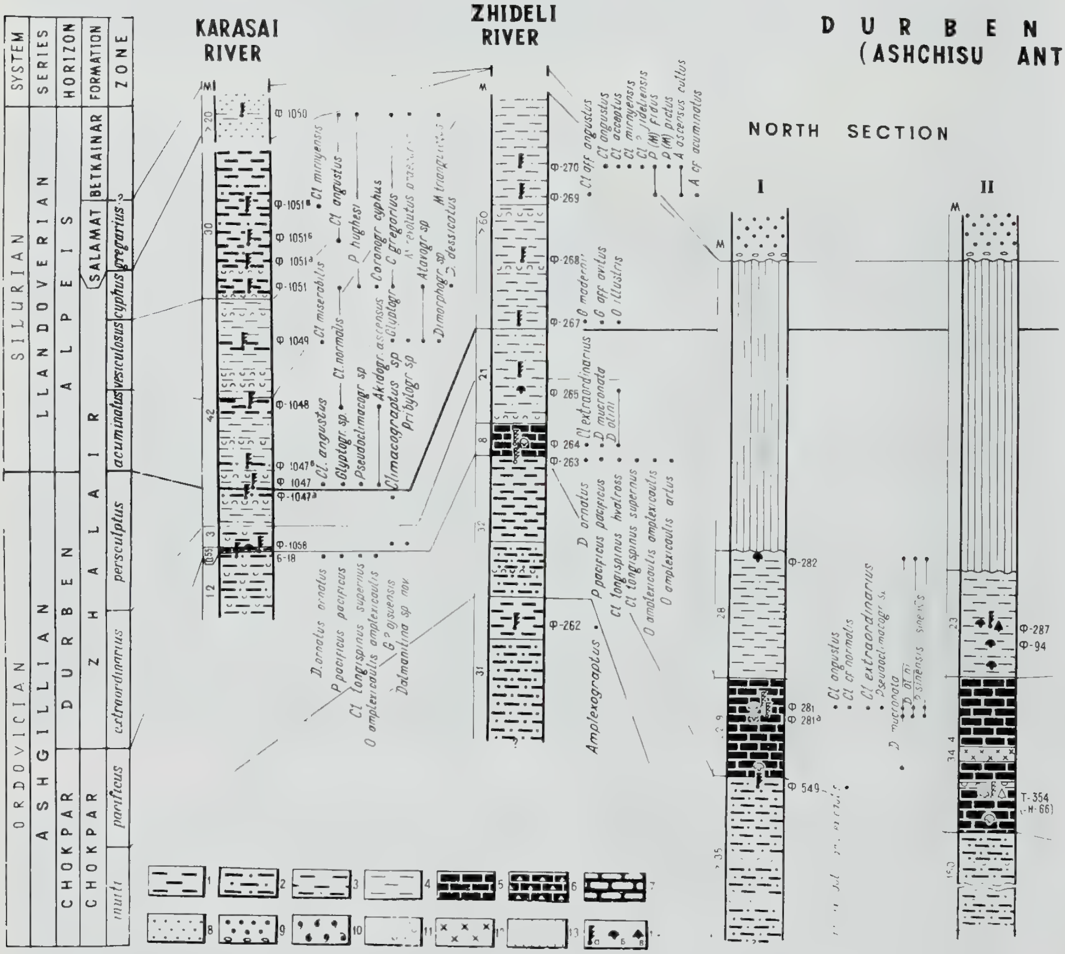
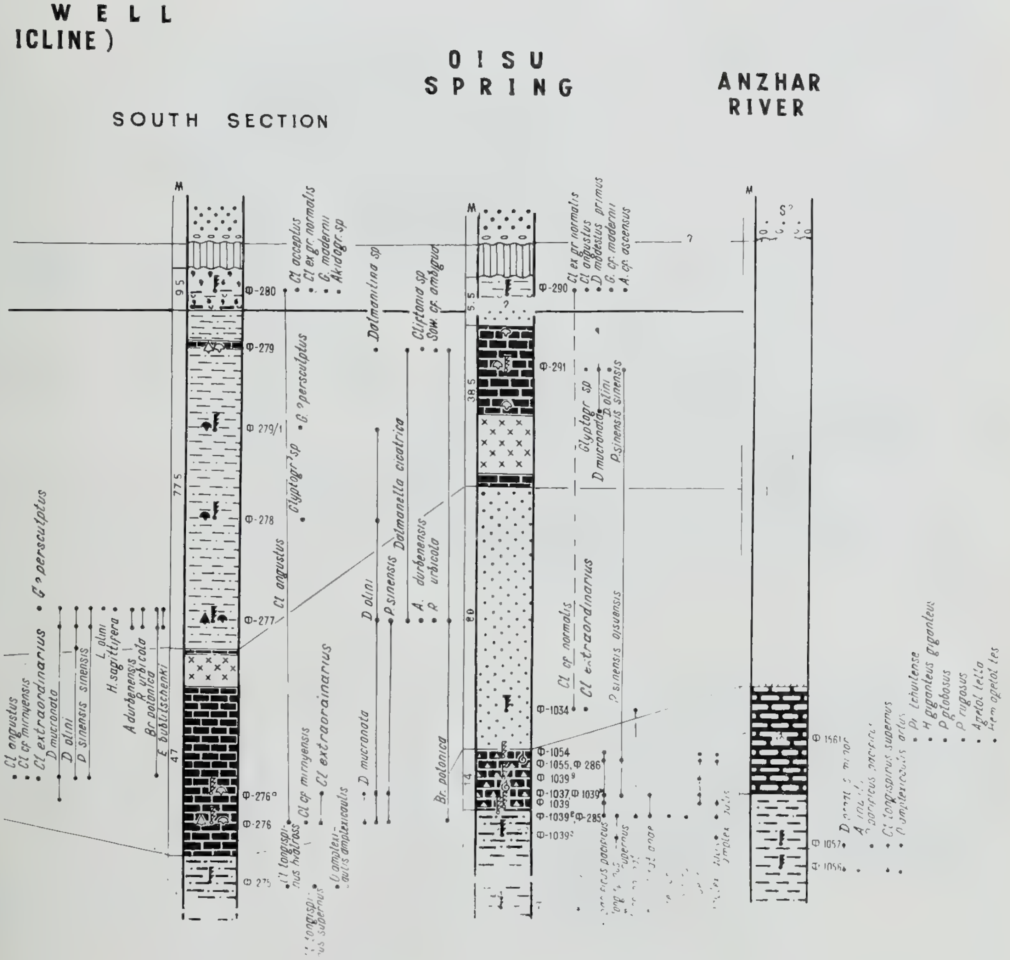
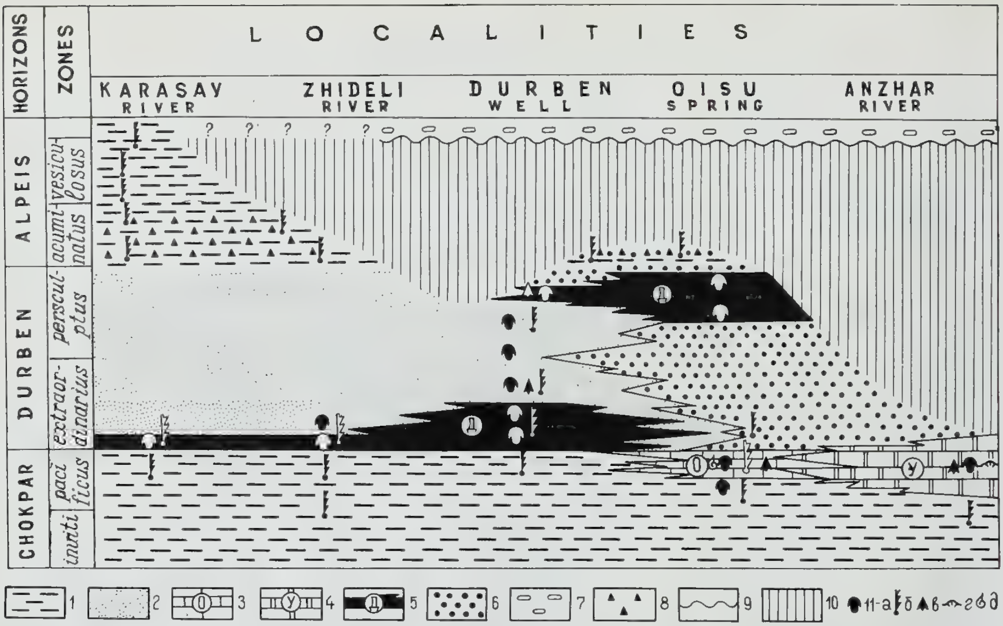

SYSTEM
SERIES
HORIZON
FORMATION
ZONE
SILURIAN
LLANDOVERIAN
ALPEIS
BETKAINAR
SALAMAT
ZHALAIR
gregarius
cyphus
vesiculosus
acuminatus
persculptus
extraordinarius
pacificus
ORDOVICIAN
ASHGILLIAN
DURBEN
CHOKPAR
KARASAI RIVER
ZHIDELI RIVER
DURBEN (ASHCHISU ANT
NORTH SECTION
I
II
Ф 1050
Ф-1051ᵃ
Ф 1051ᵇ
Ф 1051ᵃ
Ф-1051
Ф 1049
Ф-1048
Ф 1047ᵇ
Ф 1047
Ф-1047ᵃ
Ф-1058
6-18
Ф-270
Ф-269
Ф-268
Ф-267
Ф 265
Ф 264
Ф-263
Ф-262
Ф-282
Ф 281
Ф 281ᵃ
Ф 549
Ф-287
Ф-94
T-354
Cl. miseratilis
Cl. mirnyensis
Cl. angustus
Cl. normalis
P. hughesi
Coronogr. cyphus
C. gregarius
Glyptogr. sp.
Akidogr. ascensus
Climacograptus sp.
Pseudoclimacogr. sp.
Pribylogr. sp.
Atavogr. sp.
Dimorphogr. sp.
M. triangulatus
D. ornatus
P. pacificus pacificus
Cl. longispinus supernus
O. amplexicaulis amplexicaulis
G.? ojsuensis
Dalmanitina sp. nov.
Cl. aff. angustus
Cl. angustus
Cl. acceptus
Cl. mirnyensis
Cl.? zhidelensis
P. (M.) fidus
P. (M.) pictus
A. ascensus cultus
A. cf. acuminatus
D. modestus
G. aff. avitus
O. illustris
Cl. extraordinarius
D. mucronata
D. ultimus
D. ornatus
P. pacificus pacificus
Cl. longispinus hvalross
Cl. longispinus supernus
O. amplexicaulis amplexicaulis
O. amplexicaulis arcus
Amplexograptus
Cl. angustus
Cl. cf. normalis
Cl. extraordinarius
Pseudoclimacogr. sp.
D. mucronata
D. ornatus
D. sinensis
1
2
3
4
5
6
7
8
9
10
11
12
13
14

Fig. 4 Chart showing a correlation of the Ordovician–Silurian boundary deposits in the Chu-Ili Mountains. 1, black mudstones, siltstone and silty mudstones; 2, dark grey to black mudstone and siltstone; 3, grey tuffaceous pelite, tuffaceous mudstone; 4, grey siltstone, mudstone, fine-grained sandstone; 5, dark grey fine-crystalline evenly bedded limestone, sometimes clayey; 6, detrital


WELL
ICLINE)
SOUTH SECTION
OISU SPRING
ANZHAR RIVER
M
9.5
77.5
47
Ф-280
Ф-279
Ф 279/1
Ф-278
Ф-277
Ф-276а
Ф-276
Ф 275
Cl. acceptus
Cl. ex gr. normalis
G. madernii
Akidogr. sp
Glyptogr. ? sp
G. ? persculptus
Cl. angustus
Cl. cf. mirnyensis
Cl. extraordinarius
Cl. longispinus hvalross
Dolmanellina sp
Cliftonia sp
Sow. cf. ambigua
Dolmanella cicatrica
D. olini
P. sinensis
A. durbenensis
R. urbicola
D. mucronata
Br. polonica
O. amplexicaulis
Cl. angustus
Cl. cf. mirnyensis
Cl. extraordinarius
D. mucronata
D. olini
D. sinensis sinensis
L. olini
H. sagittifera
A. durbenensis
R. urbicola
Br. polonica
E. bublitschenki
G. ? persculptus
M
5.5
38.5
80
14
Ф-290
Ф-291
Ф-1034
Ф-1054
Ф-1055, Ф 286
Ф 1039
Ф-1037, Ф 1039
Ф 1039
Ф-1039, Ф-285
Ф-1039
Cl. ex gr. normalis
Cl. angustus
D. modestus primus
G. cf. madernii
A. cf. ascensus
Glyptogr. sp
D. mucronata
D. olini
P. sinensis sinensis
Cl. cf. normalis
Cl. extraordinarius
P. sinensis oisuensis
?
M
S?
Ф 1561
Ф 1057
Ф 1056
Pr. tenuilense
H. giganteus giganteus
P. globosus
P. rugosus
Agetol. tetia
Hem. agetolites
Cl. longispinus supernus
O. amplexicaulis

limestone; 7, bioherm limestone; 8, middle and coarse-grained polymictic sandstone; 9, conglomerates and coarse-grained polymictic sandstones; 10, tuffaceous sandstone; 11, fine-clastic acid tuff and tuffite; 12, diorite sill; 13, non-deposition; 14, fossils: (a) graptolites, (b) trilobites, (c) brachiopods.


HORIZONS
ZONES
LOCALITIES
KARASAY RIVER
ZHIDELI RIVER
DURBEN WELL
OISU SPRING
ANZHAR RIVER
ALPEIS
DURBEN
CHOKPAR
acuminatus
vesiculosus
persculptus
extraordinarius
pacificus
inuriti
?
Д
О
У
1
2
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4
5
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б
в
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д

Fig. 5 Chart showing the lateral variation of different lithogenetic types within the Ordovician–Silurian boundary interval in South Kazakhstan. D, Durben Limestones; O, Ojsu Limestones; U, Ulkuntas Limestones. 1, black mudstones; 2, green sandstone; 3, detrital thin-bedded microcrystalline limestone; 6, sandstone; 7, conglomerate and gritstone; 8, tuffite; 9, unconformity; 10, non-deposition; 11, fossils: (a) trilobites, (b) graptolites, (c) brachiopods, (d) corals, (e) other fauna groups.

the *acuminatus* graptolite assemblage may be due to the widespread early Llandovery transgression. A great crisis in graptolite evolution within the *extraordinarius* and *persculptus* Zones took place at the end of the Ordovician regressive cycle.

The basal lower Silurian deposits (the *acuminatus* Zone) outside the Chu-Ili Mountains are established in eastern Central Kazakhstan in the Otyzbes Mountains, near the Kombabasor Lake east of the town of Bajanaul and at the watershed of the Akzhar–Zhartas Rivers northeast of Karaganda (Bandaletov 1969; Apollonov *et al.* 1980; Fig. 1 herein).

The uppermost Ashgill deposits (the *Dalmanitina mucronata* beds of the Durben horizon) are known from the Zhaksykon River basin at the Sarysu–Teniz watershed in the Chingiz Range (near the town of Akdombak) and south-western Chingiz area (Nikitin 1972; Nikitin *et al.* 1980). The systemic boundary in the regions within the neritic development is defined by the appearance of the diagnostic brachiopods *Eospirifer cinghizicus* and *Holorhynchus cinghizicus* and tabulate corals (Borisyak *et al.* 1969; Nikitin 1972).

However, direct correlation between the graptolite and shelly sequences within the Silurian basal beds is still not fully established, and the problem of the identification of shelly faunas diagnostic of the *acuminatus* Zone remains open in Kazakhstan as elsewhere.

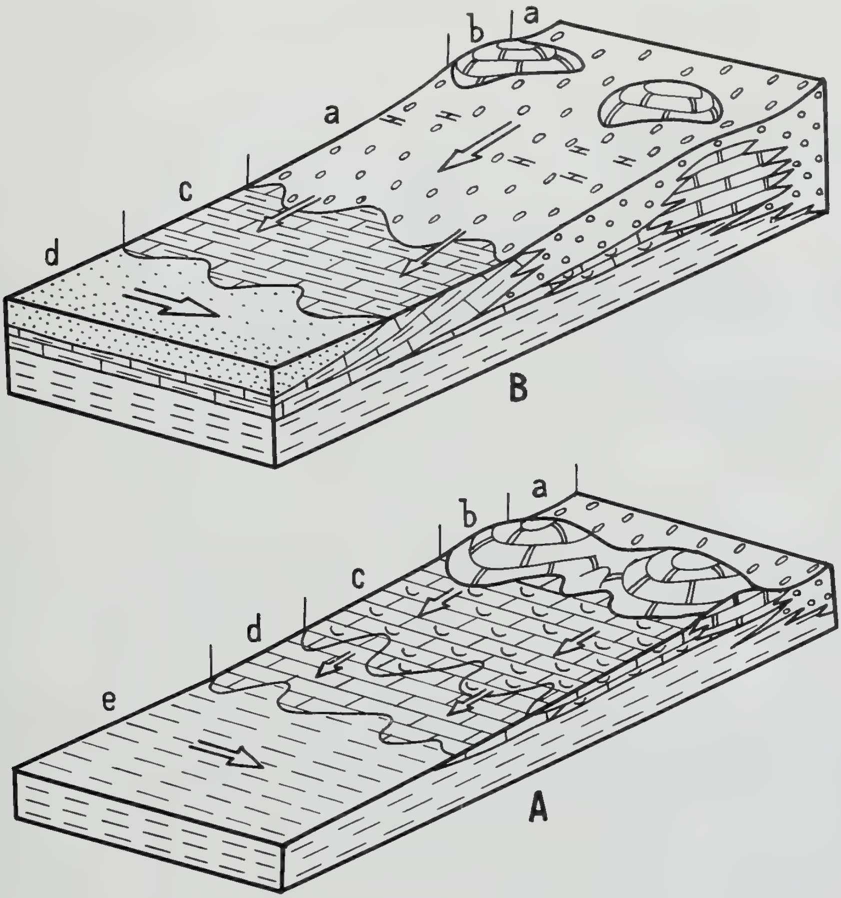

Fig. 6 Schematic depositional patterns in the South Kazakhstan Palaeo-basin in the uppermost Ordovician. A (*supernus* Zone): a, coarse sandstones; b, biohermal (Ulkuntas) limestones; c, detrital (Ojsu) limestone; d, microitic (Ojsu) limestone; e, black (Chokpar) mudstones.
B (*extraordinarius* and *persculptus* Zones): a, coarse sands; b, biohermal limestones; c, thin-bedded micritic (Durben) limestones with *Dalmanitina* association; d, fine sandstones with *Dalmanitina–Hirnantia* association.
Arrows indicate the direction of transport of the clastic material.

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The Ordovician–Silurian boundary in Saudi Arabia

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Synopsis

An account is given of the environments of deposition across the Ordovician–Silurian boundary which occurs within the Tabuk Formation, Saudi Arabia. The results of much recent work are appraised and earlier conclusions are reassessed with respect to it. The late Ordovician glaciation is considered to have been a prime factor influencing sedimentation, for example by restricting land derived clastic input. There appears to be no regionally significant depositional hiatus, and the beds about the boundary are best regarded as conformable. The general environment of deposition was of prograding sandstones, tidal flats, delta cycles and intermittent marine incursions on a tectonically stable structural platform. A basically normal graptolite sequence is deduced across the boundary region, and a précis is given of the relative dating achieved by these and other fossil groups.

Introduction

Early Palaeozoic rocks of the Arabian Peninsula are almost exclusively siliciclastics whose primary source was erosion from the western part of the Precambrian Arabian Shield. These rocks were successively deposited to the east along a regressive shoreline in fluvio-deltaic and shallow water marine environments. The Ordovician–Silurian boundary in Saudi Arabia occurs within the Tabuk Formation of this suite of rocks.

The Tabuk Formation was originally designated by R. A. Bramkamp in 1954 in an unpublished report of the Arabian American Oil Company. His definition in amended form was presented on U. S. Geological Survey Miscellaneous Geologic Investigations Map I-270A (1963). The formation was formally defined by Powers *et al.* (1966). A summary of details of the formation is given by Powers (1968).

The type section, in the Tabuk area of northwest Saudi Arabia (Fig. 1), consists of 1071 m of shale, siltstone and sandstone, deposited in shallow water in a complex of fluviatile, littoral beach, deltaic, and tidal flat sediments. Holomarine shale members, recording marine transgressive phases, occur at the base, near the middle and near the top. These are designated, respectively, the Hanadir, the Ra‘an, and the Qusaiba shales. However, in the vicinity of the type section, only the basal member, the Hanadir, shows easily mappable lateral continuity.

Powers (1966) designated a reference section of 677·2 m thickness in the Qasim (Qusaiba) area (Fig. 1) which is a composite section from several excellent exposures in the vicinities of Jebal Hanadir, Khashm Ra‘an, and in the Qusaiba depression. For the local definition of the Ordovician–Silurian boundary this section is best, with the advantages that (1) all three holomarine shale members are well developed and well exposed, (2) all three shale members are graptolite-bearing, (3) additional fossil collections, including graptolites and trilobites, have been made in more recent years and serve to refine previous age assignments and stratigraphical relationships within the formation on outcrop as well as in subsurface areas several hundred kilometres to the east, and (4) glacial beds recording an ‘end-of-the-Ordovician’ glaciation event and stratigraphical and sedimentary details have been recently studied in the area. Fig. 2 shows a generalized stratigraphical section in the Qasim area.

Stratigraphy and sedimentation

Rocks of the Tabuk Formation were deposited in shallow water on a very broad and extensive, gently sloping epicontinental shelf, reflecting the underlying basement structure of a nearly flat, gently dipping, stable homoclinal platform (Powers *et al.* 1966). Present dips on outcrop average less than 2° eastward and have been little disturbed since deposition. Graptolitic shales

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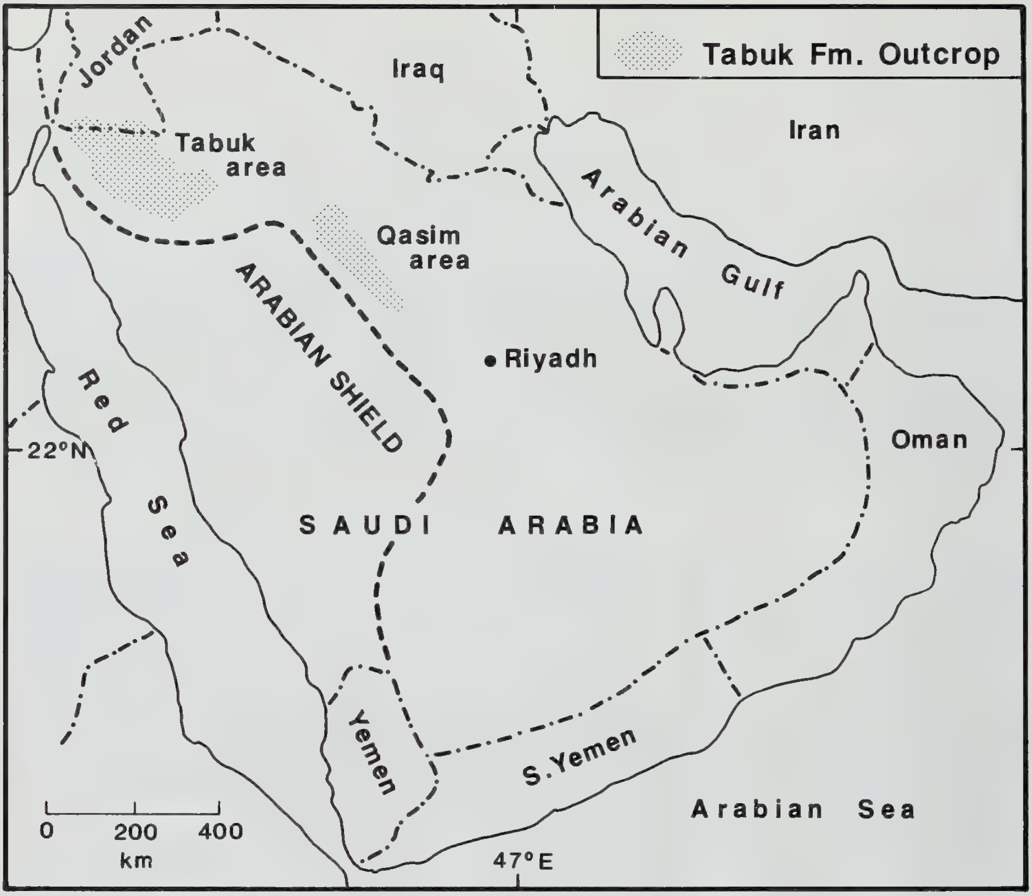


Fig. 1 Outcrop map of the Tabuk Formation in Saudi Arabia. Equivalent rocks on the surface and in the subsurface have been found as far east as Oman.

and sands with other macrofossils and trace fossils as well as a palynomorph suite of chitinozoans, acritarchs and plant spores occur on surface outcrop as well as in the deeper subsurface section of the eastern part of Saudi Arabia and Oman. The Tabuk Formation gradually thickens basinward to the east, where it is extensive in the subsurface, but, except that the three marine shales tend to be less distinct as discrete units, no gross changes in facies or depositional environment are apparent. Carbonate beds are known only as rare thin lenses at outcrop.

Lithologies of the Tabuk Formation comprise three basic types: (1) medium-grained, partly cross-bedded, partly massive-bedded, channelled sandstone, (2) fine-grained, laminated and ripple-marked, micaceous sandstones and siltstones, and (3) laminated and micaceous, fossiliferous shales, the Hanadir, the Ra‘an and the Qusaiba. The shales grade upwards through siltstone interbeds at the top into type 2 lithology. Tabuk lithologies are thus arranged in three generally coarsening-upward cycles that, together with the regional sedimentological and structural framework, suggest deposition in a prograding deltaic environment dominated by fluvial sediment input. Lithology type 1 probably represents material derived from a fluvially-fed delta plain and deposited in the distributary system of a delta front; type 2, fine sandstone and siltstone, may have been deposited in intermediate mouth bars; and type 3 is considered to represent a mud-dominated platform in the pro-delta, off-shore area, where holomarine conditions prevailed. Each of the three cycles from bottom to top probably represents sand and silt facies of a delta plain and delta front prograding during periods of eustatic stand-stills over pro-delta muds, which were the product of intermittent, possibly eustatically controlled, marine incursions.

Intertidal deposition as part of the delta plain appears to have been widespread, *Skolithos* beds and tidalite sands being prominent towards the top of the Ordovician part of the Tabuk Formation on outcrop as well as in the subsurface. A non-barred, tidally dominated, sandy coastline was probably present, where extensive fluvially-dumped sediments were contemporaneously reworked and redistributed during periods of active progradation.

Graptolite zonations, documentation of the glacial event, and sedimentary observations are the principal aids available in the area to define the nature of the Ordovician–Silurian boundary. Analysis of the Ra'an and Qusaiba shales and the intervening sandstone is particularly informative, since these units bracket the boundary. (The Hanadir, the basal shale member of the Tabuk, of Llanvirn age, is not discussed here, except briefly in the section on biostratigraphy below.)

The Ra'an is the least distinctive and persistent of the three shale members. At the type locality at Khashm Ra'an (latitude 26° 52′ N, longitude 43° 23′ E), the lithology consists of 67 m of green-grey, silty, micaceous shale and fine-grained, red-brown, ripple marked, micaceous sandstone and siltstone with trace fossils towards the top. Glacial beds are erratically associated with the top of the Ra'an in many places at outcrop. Very rare graptolites, trilobites, brachiopods and molluscs are concentrated in several thin zones at the bottom and top.

The range of the graptolite *Orthograptus amplexicaulis*, which occurs in the lower part of the Ra'an, is from the *clingani* Zone to the *anceps* Zone. *Glyptograptus persculptus* occurs at the top of the Ra'an and, although formerly considered to represent the lowest Silurian, is now taken as uppermost Ordovician. The trilobites indicate a less precise age ranging from about the middle Caradoc to about the late Ashgill. Overall considerations indicate the Ra'an member at outcrop is probably late Caradoc to the latest Ashgill, *persculptus* Zone, in age.

The sandstone overlying the Ra'an, which is similar to the sandstone underlying it, is partly cross-bedded, partly massive-bedded, medium-grained and occasionally channelled. This unit, about 240 m thick in the Qasim area, is probably lower Rhuddanian in age because of its conformable position below the well-dated Aeronian Qusaiba shale and above the *persculptus* Zone. It is generally barren of body fossils, but poorly preserved moulds of molluscs and brachiopods (mostly lingulids) are sometimes present. Trace fossils such as *Cruziana* are frequent.

The Qusaiba shale, like the Ra'an, is erratically distributed along the length of the outcrop. At its best exposure and type locality in the Qusaiba Depression (26° 53′ N, 43° 34′ E), it consists of 57 m of varicoloured, red and grey-green laminated shale with thin interbeds of yellow shale, and red, hematitic, ripple-marked, micaceous and fine-grained sandstone with trace fossils towards the top. A medium-grained, cross-bedded sandstone overlies the shaly-silty interbedded unit. The Qusaiba is especially rich in graptolites, but also contains rare trilobites, brachiopods and molluscs. Graptolites serve to date the Qusaiba as Aeronian, *convolutus* Zone.

Nature of the Ordovician–Silurian boundary

In the Arabian section, both on outcrop and in the subsurface to the east, the *persculptus* Zone is present near the top of the Ra'an shale. On the surface, *persculptus* occurs just below the glacial beds. While this zone has not been documented above the glacial horizon on outcrop, in the subsurface it occurs just above a diamictite suspected of being of glacial origin (Fig. 2).

The contact between the Ordovician and the Silurian, both at outcrop and in the subsurface further to the east, is apparently conformable. Nothing appears to have happened across the boundary that drastically altered the depositional mode characteristic throughout the Tabuk Formation of prograding sands, delta cycles, and intermittent marine incursions on a tectonically stable structural platform. Within the Ra'an, however, extreme cold water conditions were apparently manifested in an impoverished fauna, and local glacial activity took place within the top part of the Ra'an. Fluvioglacial channels, tillite deposits, striated and faceted megaclasts, exotic igneous boulders, pro-glacial sandstone, and other evidence of glaciation occur in this part of the section (McClure 1978; Young 1981). This event is assumed to be approximately coeval with glaciation at this time in north Africa (Beuf *et al.* 1969; Hambrey & Harland 1981).

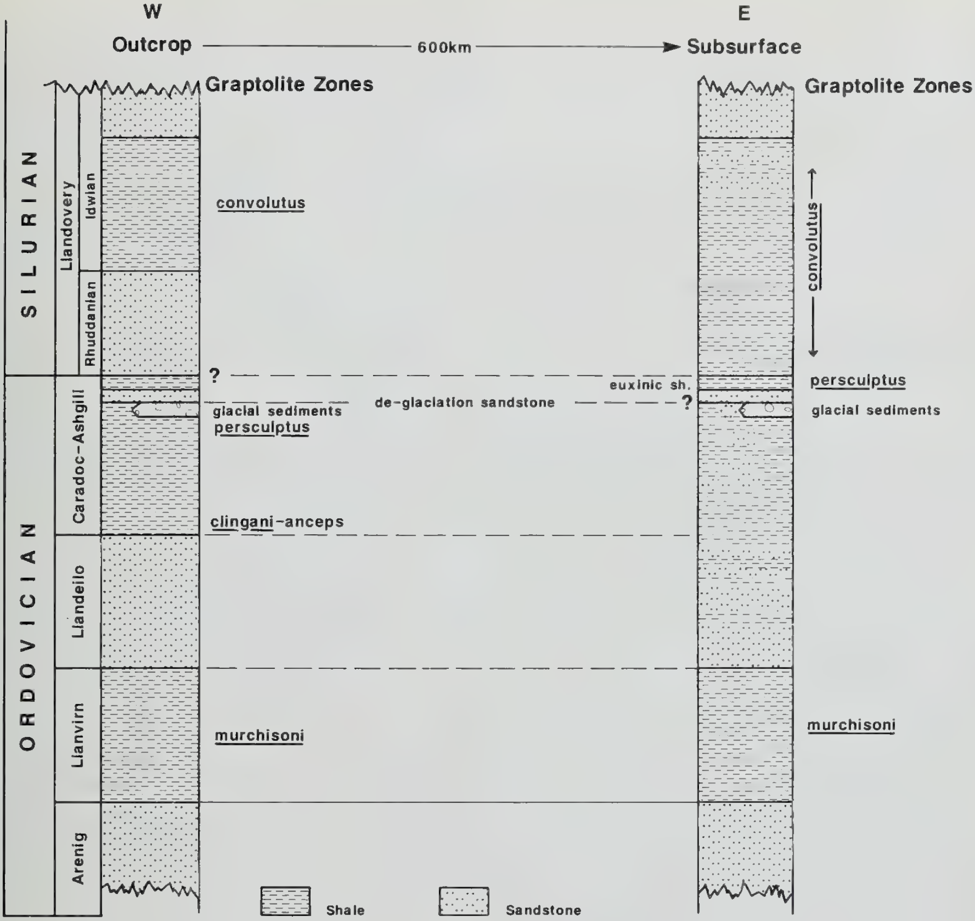
The glaciation in Saudi Arabia is confined within the top part of the Ra'an, apparently within the *persculptus* Zone, but is ice-marginal and ice-contact and not glacio-marine. Subaerial exposure due to sea level drop at the maximum of glaciation during the later phase of the Ashgill probably occurred. Thus, super-imposed upon the Ra'an is a subsidiary sequence of events composed of (1) glacio-eustatic sea level regression, during which glaciation took place, (2) deglaciation during which glaciofluvial and fluvial sands were deposited, finally followed by (3) glacio-eustatic sea level rise, during which the upper part of the *persculptus* Zone shale was deposited. Sea level dropped again toward the beginning of the Silurian and regressive sands were deposited in Rhuddanian time. In later Llandovery (Aeronian) time, a marine transgression apparently unrelated to glacial events deposited the Qusaiba shale. The glacio-eustatically controlled regressive–transgressive sequence at the top of the Ra'an may be synchrononous with similar world-wide events such as those documented by Brenchley & Newall (1980) at the end of the Ordovician in the Oslo region, Norway, and those proposed by Berry & Boucot (1973). The Ordovician–Silurian boundary in Saudi Arabia may thus be taken at the base of the sandstone unit between the Ra'an and Qusaiba shales, or above the *persculptus* Zone.

Lithofacies to the east in the deep subsurface vary little from the outcrop sequence, except that the Ra'an as a discrete shale unit with easily determined top and base is not always present and the sandstone of presumed Rhuddanian age between the Ra'an and the Qusaiba at outcrop is poorly developed. The contact between the Ra'an and the Qusaiba is consequently indistinct, and the Qusaiba sequence is considerably thicker. A distinctive feature of the subsurface is a regionally persistent and prominent, thin, highly organic, pyritic euxinic black shale, often bearing common or abundant *Glyptograptus persculptus* with no benthic fossils and overlying a sandstone with diamictite suspected of being equivalent to the glacial tillite of outcrop. This shale may be equivalent to the post-glaciation upper part of the *persculptus* Zone of outcrop mentioned above and helps place the glaciation event as within the *persculptus* Zone.

The graptolite succession of the deep subsurface requires further study, but appears similar to that of the outcrop. Several differences are that graptolites assignable to the *clingani* to *anceps* Zones as found at the base of the Ra'an on outcrop have not been documented in the subsurface, and a graptolite suite assignable approximately to the boundary between the *magnus* and *leptotheca* Subzones of the *gregarius* Zone has been recovered in one drill hole. The most perplexing anomaly, however, is that, in another representative drill hole with continuous core sequence, a *convolutus* Zone graptolite suite occurs within about 6 m of the euxinic *persculptus* Zone shale. Intervening graptolite zones of the lower Llandovery therefore appear to be largely missing or drastically telescoped. (See Note, p. 163).

The 'missing' graptolite zones are assumed to be represented on outcrop by the non-fossiliferous Rhuddanian sandstone and their apparent absence in the deeper section where this sandstone is not present and shales were continuously deposited is puzzling. However, these zones are also 'compressed' in some standard British successions (R. B. Rickards, personal communication) and lower Llandovery marine fossils are rare on a worldwide basis (Berry & Boucot 1973). The apparent gap in the graptolite succession of Saudi Arabia is probably not due to events peculiar to the Arabian platform. It is tempting to consider the euxinic black shale as well as the condensing or absence of the early Llandovery graptolite zones as in some way related to the glaciation event. Cessation or drastic restriction of fluvial flow regimes and consequent constriction of clastic input due to tie-up of water in glacial ice may have resulted in stagnant, euxinic conditions in more distal intra-platform areas on what was already a cold water, carbonate-starved platform. Fig. 2 presents outcrop and subsurface correlations within the Tabuk Formation.

Thus, a firmer calibration of a time scale with depositional and climatic events and faunal occurrences is essential to define more precisely the nature of the Ordovician–Silurian boundary on the Arabian platform and correlate it with sequences elsewhere. The evidence accumulated to date, however, is informative, and the following conclusions can be tentatively made.


W
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Outcrop
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Graptolite Zones
SILURIAN
ORDOVICIAN
Llandovery
Idwian
Rhuddanian
Caradoc–Ashgill
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Arenig
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persculptus
glacial sediments
de-glaciation sandstone
?
glacial sediments
persculptus
clingani–anceps
murchisoni
murchisoni
Shale
Sandstone

Fig. 2 Section comparing the Tabuk Formation in the outcrop of NE Saudi Arabia (left) with that in the subsurface to the east (right). The three shale horizons at outcrop are termed Hanadir (Llanvirn), Ra'an (Caradoc to basal Llandovery) and Qusaiba (Middle Llandovery). The Idwian Stage shown is now regarded as the lower part of the Aeronian Stage. The base of the Tabuk Formation is at the base of the Llanvirn.

1. Rates of sediment deposition may have varied on the Arabian platform across the Ordovician–Silurian boundary and can probably be attributed to the effects of glaciation. Land-derived clastic input may have been drastically restricted, resulting in euxinic, starved and stagnant areas, but:
2. No regionally significant depositional hiatus is evident and the contact between the Ordovician and Silurian may be considered conformable.
3. Nothing except glaciation happened at the boundary to upset significantly the mode of deposition characteristic throughout the Tabuk Formation of prograding sandstone, tidal flats, delta cycles and intermittent marine incursions on a tectonically stable structural platform.
4. The graptolite succession across the boundary appears normal, the apparent gap of early Llandovery graptolite zones being probably attributable to world-wide events and not intra-platform activity.
5. The significant boundary event on the Arabian platform appears to have been the glaciation at the end of the Ordovician that affected sedimentation rates and faunal suites.

Biostratigraphy

Early fossil collections listed by Powers *et al.* (1966) and Powers (1968) were sparse. Additional surface collecting in more recent years in the Qasim (Qusaiba) area has provided fossils that serve to refine previous age assignments as well as to reveal more about the palaeobiology of the Tabuk Formation and faunal events across the Ordovician–Silurian boundary. Drill hole cores available in recent years from the deep subsurface of the eastern part of Saudi Arabia, where rocks across the boundary are extensively distributed, also provide useful information.

All three shale members of the Tabuk Formation, the Hanadir, the Ra'an, and the Qusaiba, are fossiliferous holomarine shales deposited as pro-delta muds. Intervening sands are largely of tidalite and shoreface origin and are mostly unfossiliferous of body fossils, but frequently contain trace fossils including *Skolithos* and *Cruziana.* Thin siltstone beds near the tops of the shales rarely contain poorly preserved moulds of bivalves, brachiopods (lingulids) and trilobites. All three shale units contain graptolites, trilobites, and an assortment of benthic fauna in addition to palynomorphs (chitinozoans, acritarchs, and spores).

Except for graptolites, trilobites, and palynomorphs, the fossil suite has been little studied. R. B. Rickards has been working with the graptolites in recent years; Thomas (1977), Fortey & Morris (1982) and El-Khayal & Romano (1985) have studied some of the trilobites, H. A. McClure is working on chitinozoan and acritarch suites and J. Gray, A. J. Boucot and H. A. McClure are currently investigating spores of possible land plant affinity. The following analysis should be considered preliminary. The following lists are comprehensive compilations from both outcrop sequences (Qasim area) and cored holes of the subsurface to the east.

Though not strictly pertinent to the boundary problem, the fossils of the Hanadir shale at the base of the Tabuk Formation are also listed. The Hanadir at its type section (26° 27′ N, 43° 27′ E) consists of 60 m of varicoloured, laminated, micaceous shale, with thin, red-brown, ripple marked siltstone and fine grained sandstone at the top. Fossils of the Hanadir include:

Graptolites: *Didymograptus murchisoni murchisoni* (Beck), *D. murchisoni geminus* (Hisinger), *D. pakrianus* Jaanusson, *D.* aff. *D. acutus* Ekstrom, *Amplexograptus* cf. *A. coelatus* (Lapworth), *A.* sp. Trilobites: *Neseuretus tristani* (Desmarest), *Plaesiacomia vacuvertis* Thomas, ?*Marrolithus* sp. Cephalopod: *Orthoceras* sp. Brachiopods: ?*Monobolina* sp. and other articulate species and unidentified lingulids. Molluscs: ?*Glyptarca* sp., unidentified bivalves, unidentified gastropods. Beyrichids and other unidentified ostracodes; unidentified conodonts and palynomorphs (chitinozoans, acritarchs, spores, and scolecodonts). Based mainly on the graptolites, the Hanadir is Llanvirn in age, *murchisoni* Zone.

The Ra'an shale contains the following fossils, derived mainly from several thin zones at the base and toward the top from outcrop and from cores of the subsurface: Graptolites: *Glyptograptus persculptus* (Salter) *s.s.*, *Orthograptus amplexicaulis* Hall *s.s.*, *Orthograptus* sp. nov., *Diplograptus* sp., *Climacograptus angustus/normalis*, ?*Dictyonema* sp., ?*Climacograptus miserabilis* and ?*Diplograptus modestus.* Trilobites: *Kloucekia* sp. and *Onnia* sp. Brachiopods: *Comatopoma* sp. or *Hirnantia* sp., others (mostly lingulids). Molluscs: unidentified gastropods and bivalves and the cephalopod *Orthoceras* sp.; unidentified conodonts and palynomorphs (chitinozoans, acritarchs, spores, and scolecodonts).

The range of *Orthograptus amplexicaulis* is from the *clingani* to the *anceps* Zones. *Glyptograptus persculptus* places the top part of the Ra'an in the *persculptus* Zone.

The Qusaiba shale contains the following: Graptolites, Suite 1: *Climacograptus scalaris* (Hisinger), *C.* aff. *C. rectangularis* Törnquist, *Glyptograptus* aff. *G. incertus* Elles & Wood, *G. tamariscus tamariscus* (Nicholson), *G.* (*Pseudoglyptograptus*) sp., *Lagarograptus* sp., *Monograptus capis* Hutt, *M. communis* Lapworth, *M. convolutus* (Hisinger), *M. decipiens* Törnquist, *M. gregarius gregarius* Lapworth, *M. lobiferus* (M'Coy), *M.* cf. *M. delicatulus* (Elles & Wood), *M.* ex gr. *tenuis* (Portlock), *Orthograptus cyperoides* Törnquist, *Petalograptus ovatoelongatus* (Kurk), *P.* sp., *Pristiograptus regularis* (Törnquist), *Pseudoclimacograptus* (*Clinoclimacograptus*) *retroversus* Bulman & Rickards, *P.* (*Pseudoclimacograptus*) sp. nov., *Rastrites spina* Richter, *Retiolites perlatus* (Nicholson), *Rhaphidograptus tornquisti* Elles & Wood. Graptolites, Suite 2: *Climacograptus tamariscus s.l.*, *Coronograptus gregarius* cf. *C. minisculus* Obut & Sobolovskaya, *Climacograptus* cf. *C. rectangularis*, *Diplograptus* cf. *D. magnus*, *Monograptus lobiferus* (M'Coy), *Pristiograptus ?concinnus*. Trilobite: *Platycoryphe dyaulax* Thomas. Bivalves: *Nuculites*, among others. Bellerophontids, unidentified gastropods; the cephalopod *Orthoceras* sp.; brachiopods: '*Camarotoechia*' and other unidentified articulates. Unidentified conodonts and palynomorphs (chitinozoans, acritarchs, spores, and scolecodonts); ostracodes, *Tentaculites*, ophiuroids and fish remains.

On graptolite evidence of Suite 1, the outcrop of the Qusaiba is Llandovery, Aeronian Stage, *convolutus* Zone. A slightly older zone in the subsurface is represented by Suite 2, from the *gregarius* Zone, approximately on the boundary between the *magnus* and *leptotheca* Subzones, still within the Aeronian.

Palaeoecology and Palaeobiogeography

The fossil content of the Tabuk Formation was the product of a remarkably stable environment and static physical conditions for a considerable period of time. Two kinds of faunal association are represented in the Tabuk suite: (1) planktic, with graptolites, chitinozoans and acritarchs, and (2) level-bottom benthic, with an epifauna of what were probably mostly vagrant shelly benthos such as brachiopods, molluscs, trilobites and ostracodes. Other taxa such as conodonts, scolecodonts and ophiuroids are also represented. Fine layering and lamination and lack of bioturbation of the shales indicates that a significant infauna was probably not present. In general terms, population densities were high for the planktic level and low for the benthic. Diversity was moderately high for the graptolites and very high for the chitinozoans and acritarchs. Shelly benthics identified to date indicate a low diversity.

Continuity of the Tabuk suite extends for hundreds of kilometres, the fossils known from cored sequences in deep drill holes in the subsurface further to the east do not differ significantly from those of outcrop. There are no obvious indications that any element of the Tabuk biota is allochthonous.

In the shales of the Tabuk, graptolites are common but of low diversity in the Hanadir, rare and of low to moderate diversity in the Ra'an, and abundant and of high diversity in the Qusaiba. Molluscs (especially bivalves), brachiopods, trilobites and ostracodes are the next most common taxa, occurring in about equal abundance and approximately equal diversities. The shelly benthos is certainly not brachiopod-dominated as in more northern biogeographic realms. Conodonts occur in all three shale members, but are very rare and to date very little is known about them. Scolecodonts occur as a minor part of the palynomorph suites. Chitinozoans and acritarchs are common to abundant and of high diversity in the Hanadir, comparatively rare and of comparatively low diversity in the Ra'an and abundant and of high diversity in the Qusaiba. Spores, including tetrahedral tetrads that possibly represent early vascular land plants, are rare to common in the palynomorph suites. Ophiuroids are very rare; only several specimens of less than 0·5 cm size are recorded from the Qusaiba. *Tentaculites* occurs rarely in the Qusaiba. *Orthoceras* is rare but ubiquitous in all three shales, being more common and robust in the Qusaiba. In one limited locality, near the base of the Ra'an, it is concentrated in small planoconvex lenses of calcareous debris associated with algal nodules. (This is the closest resemblance to the *Orthoceras* limestone lenses recorded as widespread in the Silurian of north Africa by Berry & Boucot, 1973. The Arabian occurrence possibly represents a storm event.)

All the shelly benthic species are small, brachiopods and molluscs being rarely more than one centimetre in maximum dimension. Shelly specimens appear to have been weakly calcified or subjected to early dissolution. Most of the material is composed of moulds of the composite type on poorly defined bedding planes and laminae. As in the case of composite-type moulds (McAlister 1962), fine interior and external morphological features are often well preserved. Although taxonomic diversity is generally maximised in shallow marine environments (Boucot 1981), this is not the case with the Tabuk fauna. The condition of the shelly benthics of the Tabuk shales may indicate the influence of low salinity, but cold-water conditions (especially marked during the glaciation at the end of the Ordovician) was most likely the over-riding control. Fortey & Morris (1982) regard the trilobite genus *Neseuretus*, present in the Llanvirn (Hanadir) of the Tabuk, to be a reliable and sensitive indicator of inshore facies in cold water.

Planktics do not appear to have been affected by some cold, but were clearly affected by the excessive glacial cold conditions at the end of the Ordovician. An extensive platform covered with hyposaline water can exist if an adjacent continent has a river system adequate to provide a steady influx of fresh water. In such environs today, there is a low taxic diversity, and there is no reason to think that extensive river regimes of the past flowing off large land masses may not have had the potential for producing similar hyposaline environments with appropriate faunal consequences (Boucot 1981). This condition may have prevailed on the Arabian platform during Ordovician and Silurian times. Outcrop sequences of the Tabuk sands, silts, and shales are oxidized to shades of red, pink, yellow and green. However, subsurface equivalents invariably range from light grey to dark grey and black. Tidalite sands are especially rich in carbonaceous laminae, each lamina possibly representing nutrient material transported by a single tidal event. Tabuk shales in the subsurface are usually medium grey to dark grey and black, the extreme case being the black, highly radioactive, 'sooty' shale of the subsurface *persculptus* Zone.

Some of the palynological samples yield a distinct tetrahedral tetrad type of suspected land origin. This kind of evidence for vascular land plants may be recorded as early as the Llanvirn (Hanadir shale) in the Arabian Tabuk section. Berry & Boucot (1973) suggest that a black, radioactive shale at the 'base of the Silurian' in north Africa could be due to blooms of plants in mud flats and lagoons at this time. An apparently synchronous event occurs across much of north Africa and Arabia. A readily accessible and presumably useable supply of nutrients might therefore be assumed for both planktic and bottom benthics. Nutrient kind and availability may have been a significant factor in the palaeoecology of chitinozoans and acritarchs especially, and perhaps also graptolites.

Temperature is probably the most important variable affecting both plant and animal distribution of the present and continental glaciation episodes of the end of the Ordovician and Permian–Carboniferous are associated with global diversity gradients (Boucot 1981). The change in faunal composition associated with the Ordovician glaciation is now well documented (Berry 1973; Berry & Boucot 1973). A Silurian warming followed the Ordovician cold in the area and this may be reflected in taxa of the Qusaiba fossil suite being relatively more diverse and populations denser, especially planktic ones.

In summary, the Tabuk palaeogeography and palaeoenvironment was probably that of a broad pro-delta mud substratum on a shallow-water, clastic-fed, carbonate-starved, tectonically stable platform area, with sediment and high nutrient input derived from a low, rapidly eroding landmass, possibly with primitive plant cover, and transported via extensive fluvial, tidal and deltaic systems. Conditions of low salinity and cold-water temperatures probably controlled diversity and density of parts of the faunal community. Conditions may be considered to have been optimum for planktic taxa such as chitinozoans and acritarchs, favourable for graptolites, and generally unfavourable for benthics.

Lovelock *et al.* (1981) record chitinozoans and acritarchs, trace fossils, ?dalmanellid brachiopods, and the trilobite ?*Neseuretus* from Early and Middle Ordovician rocks of the Amdeh Formation of Oman. Rocks of southern Jordan of age equivalent to the Tabuk Formation are sandstones, shales and siltstones bearing graptolites, brachiopods, bivalves and gastropods, nautiloids, *Conularia* and trace fossils such as *Cruziana* and *Skolithos* (Bender 1975). Exact equivalents in these two areas to individual units of the Tabuk Formation, the precise definition of the Ordovician–Silurian boundary, and the comparison with the Tabuk fauna remain yet to be worked out.

Conclusions

Pending further study and documentation of the palaeobiology of Arabian Ordovician–Silurian fossil suites, the following conclusions are presented:

1. The fossils of all three Tabuk shales are similar in composition, diversity, population density and abundance levels and may be taken to represent one community.

2. Two trophic levels are readily identifiable: (a) planktic—consisting of graptolites, chitinozoans and acritarchs, and (b) benthic—consisting largely of vagrants on a flat-bottom mud

substratum.

3. Water temperature was probably the main environmental control on the community.
4. Salinity was possibly a secondary control on the community.
5. Nutrient material was readily available and may have played a significant role in the palaeocology.
6. An extensive pro-delta mud platform provided the main palaeogeographical control for the bulk of the Tabuk fauna; inshore sandy facies and tidal flats were less important features.
7. The main event that affected the biological community across the Ordovician–Silurian boundary was stress imposed by glaciation at the end of the Ordovician. Otherwise, the conditions that affected the community throughout deposition of the Tabuk were also operative in boundary times.
8. Similarities in the palaeobiology, palaeogeography and palaeoecology occur in the platform rocks of the north African Silurian sections.

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Note added in page proof. The *atavus* Zone (Rhuddanian) has recently been documented in the Arabian Silurian section. *Atavograptus atavus* (Jones) is present in both the Tabuk area of outcrop and the deep subsurface of eastern Saudi Arabia. In the outcrop section, it occurs with *Climacograptus normalis* Lapworth.

The Ordovician–Silurian boundary in Morocco

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Synopsis

At only one locality, Moulay bou Anane, in Jbilet, the *persculptus* and *acuminatus* Zones are both found, although the *acuminatus* Zone is known from many localities throughout Morocco. The early Llandovery usually consists of transgressive shales, ranging from *acuminatus* Zone to *cyphus* Zone and above in age, overlying usually unfossiliferous glacial sandstones and microconglomerates of the latest Ordovician, from one of which a *Hirnantia* fauna is recorded.

General survey

The Ordovician–Silurian boundary in Morocco is always marked by a very acute change of facies between the two systems. The glacial episode which concludes the Ordovician deposited relatively coarse material, such as sandstones, quartzites and microconglomeratic clays, which strongly contrast with the fine argillaceous or siliceous deposits which characterize the beginning of the Silurian. Consequently, the scenario is one of more or less important interruption in sedimentation, the development of glaciogenic sediments, and the transgressive development of a Silurian sea after the melting of the continental ice sheet. Under these conditions, the faunas of the two systems are naturally different, apart from the single exception of Jbilet, at Moulay bou Anane (Locality 1, of Fig. 1), where selected graptolites for the official boundary (Cocks 1985), *Glyptograptus persculptus* Salter and *Akidograptus acuminatus* (Nicholson), succeed each other in the same section. Elsewhere, only *A. acuminatus* dates the beginning of the Silurian above more or less terminal beds of the Hirnantian:

(1) in the western Anti-Atlas, at Aïn Oui n'Deliouine (Locality 2);

(2) in the eastern Anti-Atlas, at Tizi ou Mekhazni (Tizi Ambed) (Locality 3) and at Oued Bou-Leggou (Oued bou Oubagou) (Locality 3′);

(3) on the northern slope of the central High Atlas, at Ghogoult (Locality 4) and west of Tiwghaza (Locality 4′);

(4) in the substratum of the Plateau des Phosphates (Locality 5);

(5) in the Moroccan central massif in the Azrou area, at Bou Ourarh (Locality 6);

(6) in the Palaeozoic inliers of the north of the middle Atlas at Tazekka (Locality 7) and Immouzer du Khandar (Locality 8).

Some other outcrops of the transgressive Silurian are still later Rhuddanian:

(a) in the central Anti-Atlas, at Rich Mel'Alg, where graptolitic beds with *Cystograptus vesiculosus* (Nicholson), *Dimorphograptus*, and *Coronograptus cyphus* (Lapworth) are separated by a red layer from sandstones and clays of the Deuxième Bani (Upper Ashgill);

(b) in the coastal Meseta, at Oulad Saïd, south of Casablanca, where *Atavograptus atavus* (Jones) occurs in a boring;

(c) in the Qasbat-Tadla-Azrou area, at Jbel Eguer-Iguiguena, where the same association as in (a) occurs.

For (b) and (c) it is not possible to determine with precision the age of the underlying beds.

The very widespread Silurian in Morocco more generally begins either with Aeronian beneath a siliceous facies alternating with phthanitic ribbons, more sandy in the Anti-Atlas at the east of the meridian of Icht, or sometimes with argillaceous–siliceous Telychian, or, in rare cases, with the upper Wenlock and/or Ludlow.

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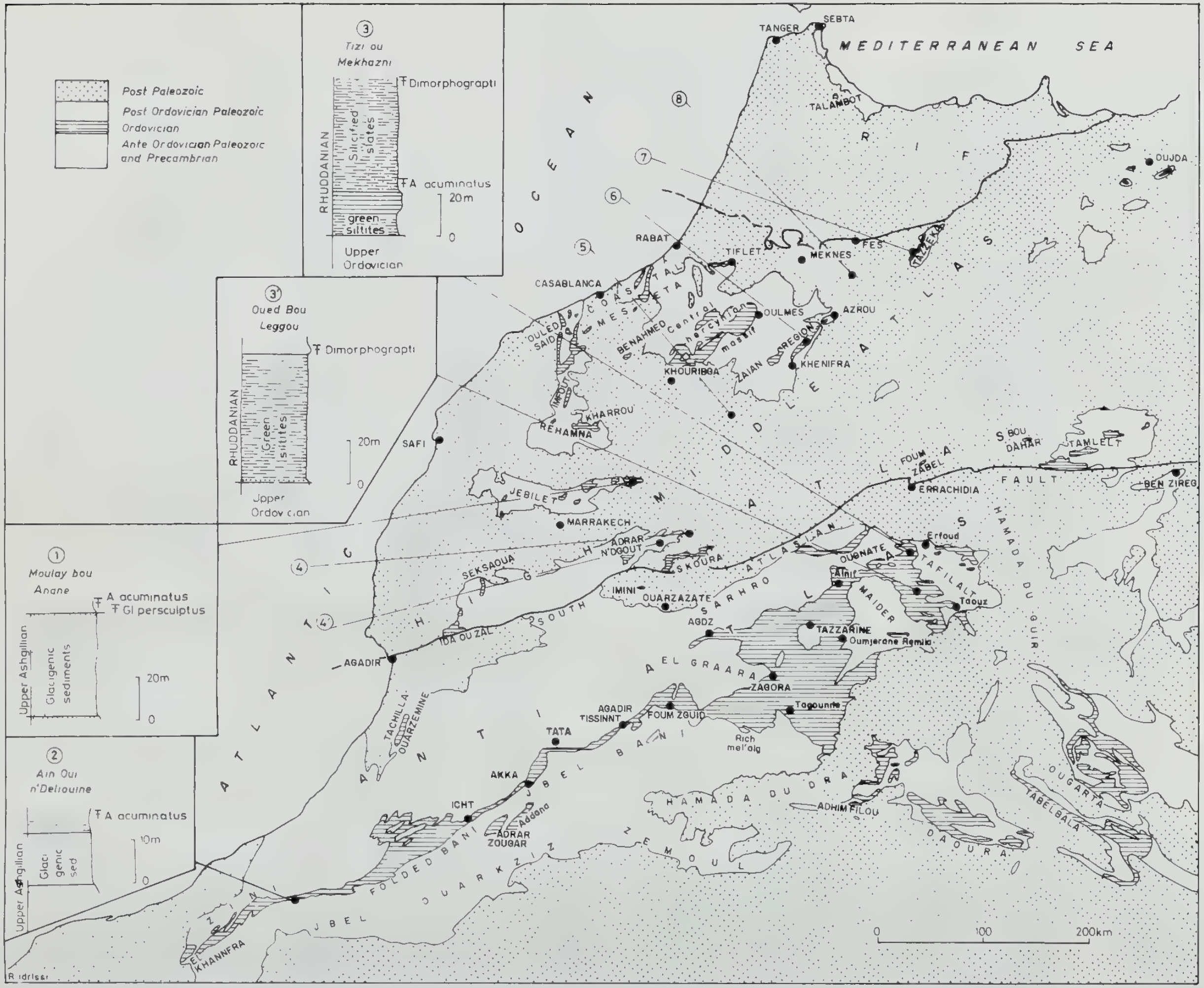
Post Paleozoic
Post Ordovician Paleozoic
Ordovician
Ante Ordovician Paleozoic and Precambrian
③ Tizi ou Mekhazni
Dimorphograpti
RHUDDANIAN
Silicified slates
A acuminatus
20m
0
green-siltites
Upper Ordovician
③' Oued Bou Leggou
Dimorphograpti
RHUDDANIAN
Green siltites
20m
0
Upper Ordovician
① Moulay bou Anane
A acuminatus
Gl persculptus
Upper Ashgillian
Glacigenic sediments
20m
0
② Ain Oui n'Delouine
A acuminatus
10m
0
Upper Ashgillian
Glaci genic sed
MEDITERRANEAN SEA
ATLANTIC OCEAN
TANGER
SEBTA
TALAMBOT
RIF
OUJDA
FES
MEKNES
TAZZEKA
RABAT
TIFLET
CASABLANCA
COASTAL MESETA
OULED SAID
BENAHMED
Central hercynian massif
OULMES
AZROU
ZAIAN REGION
KHENIFRA
KHOURIBGA
IMFOUT
KHARROU
REHAMNA
SAFI
MIDDLE ATLAS
HIGH ATLAS
JEBILET
MARRAKECH
AORAR N'DGOUT
SEKSAOUA
SKOURA
IMINI
OUARZAZATE
IDA OU ZAL
SOUTH ATLAS FAULT
AGADIR
SAHRRO
AGDZ
OUGNATE
Alnif
MAIDER
TAFILALT
Erfoud
Taouz
TAZZARINE
Oumjerane Remlia
FOUM ZABEL
ERRACHIDIA
S BOU DAHAR
TAMLELT
BEN ZIREG
HAMADA DU GUIR
EL GRAARA
ZAGORA
Tagounite
AGADIR TISSINNT
FOUM ZGUID
Rich mel'alg
TATA
AKKA
ICHT
AORAR ZOUGAR
Addana
JBEL BANI
FOLDED BANI
ANTI ATLAS
TACHILLA-OUARZEMINE
HAMADA DU DRA
ADHIM FILOU
OUGARTA
TABELBALA
DAOURA
ZEMOUL
OUARKZIZ
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0
100
200km
R. Idrissi

Fig. 1 The Ordovician outcrops of northern Morocco and ten localities with *Akidograptus acuminatus* described in the text.

Description of partial sections

(1) *Eastern Jbilet, Moulay bou Anane, Locality 1* (Topographical Sheet Attaouia ech Chaïbia, 1:50 000, at x = 322, y = 157·2) (Fig. 1). Roch (1939) described this area as forming part of the 'Mountains to the East of Marrakech'. Huvelin (1977) emphasized the peculiar features of the Hercynian massif of Jbilet. Huvelin and others refined the section near the boundary in 1980. Roch only pointed out that 'Miss G. Elles and G. Waterlot have recognised: *Monograptus* (sic) *modestus, M. sandersoni, M. cyphus, M. revolutus, Glyptograptus incertus, G. persculptus, Climacograptus Törnquisti* and *Cl. normalis* from the base of Llandovery' (p. 113). Specimens of *Glyptograptus persculptus*, determined by G. Elles, were obtained from a siliceous sandstone, weathered pink-beige, but more greyish on fresh fracture, in beds on which they are nearly orientated. They are of great size, the septum always starting at the fourth or fifth theca, and they are preserved as internal moulds, in whole or half relief.

Vertical section 1 summarizes more recent collections. The usual suite of terminal Hirnantian occurs over 20 m and consists of microconglomeratic clays, argillites, and sandstones with orientated sedimentary features. This is followed by a layer of quartzose sandstone not much different from those of Roch, but coarser, which yields dispersed *G. persculptus* with a few more irregularly orientated and smaller forms with a septum beginning at a lower level (in the third theca when visible). They are always internal moulds and are apparently narrower than those identified by G. Elles, but they show more relief. The thickness of the layer is 30 cm and it can be presumed that the Roch assemblage is rather nearer the top than the base.

Above this coarse facies, and without transitional beds, pink and pink-beige shales with a little mica and with a very fine cleavage, contain at their base: *Climacograptus normalis* (Lapworth), *C. miserabilis* Elles & Wood, *C. rectangularis* (M'Coy), *Diplograptus modestus* Lapworth, *Akidograptus ascensus* Davies and *A. acuminatus* (Nicholson). The thickness of this argillaceous level is 30 cm and occurs below alternations between more phthanitic beds and more or less siliceous clays which terminate the Rhuddanian. The boundary is therefore very sudden and with a sharp change of facies.

(2) *Western Anti-Atlas, Ain oui n'Deliouine, Locality 2* (Topographical Sheet Tiglit, 1:50 000, at x = 1076·4, y = 764·2). The boundary was figured in some detail in Destombes *et al.* (1985: 242, fig. 46). Above green microconglomeratic strata representing the glacial upper Ordovician, a red bed makes a clear transition with argillites which are very similar in colour, although a few are greener, and shows the same alteration and preservation for fossils as at Moulay bou Anane, although the cleavage is coarser. At the contact there is *C. normalis* and *D. modestus* and two metres above a single, small, aseptate specimen of *G. persculptus*, together with *C. normalis, C. transgrediens* Waern, *D. modestus*, and *A. acuminatus.*

The similarities between the two areas are striking for the early Silurian beds and, from the palaeontological point of view, the abundant *D. modestus* shows some intraspecific variations which recall Davies's (1929) considerations on the similarities of *G. persculptus* and *D. modestus*, and whether it is a case of convergence or of real relationship. Internal moulds in iron-oxides only emphasize, once again, all the pitfalls in determining deformed graptolites by comparison with material which has preserved its proteic skeleton. Finally, from these two localities, which appear to be the most characteristic of those actually known from Morocco, it is difficult to imagine any Ordovician–Silurian boundary without a break.

(3) *Eastern Anti-Atlas, Tizi ou Mekhazni* (*Tizi Ambed*), *Locality 3 and Bou Leggou* (*Oued Bou Oubagou*), *Locality 3'*. A peculiar feature of the sections near the boundary is the presence, above conglomeratic sandstones and quartzites and lenses with very coarse green and pink sandstones, of a green siltstone with a very probable hard ground between the two deposits.

(a) At Tizi ou Mekhazni (Topographic Sheet Erfoud, 1:100 000, at about x = 588·8, y = 73·8), Destombes *et al.* (1985: 257–258, figs 54 and 55) report 10 m of greenish silts followed by a black marker bed about 10 m thick of very fine silicified slates with tuff layers, followed by 75 m of fine silicified white, pink and reddish violet slates, the base of which includes *C. normalis, C. transgrediens, D. modestus, A. ascensus*?, and *A. acuminatus* in the first 5 m. The

Rhuddanian and the Aeronian continue up to the *M. sedgwickii* Zone within 125 m of siliceous sandstones, sometimes in plaquettes which weather to a very dark ferrugineous colour, but lighter on splitting.

(b) At Bou-Leggou (Topographical Sheet Erfoud, 1:100 000, at about x = 589·2, y = 56·6), the Rhuddanian includes about 60 m of green silts which contain nine levels with classic climacograptids (*Cl. normalis*, *transgrediens*, *praemedius* Waern, *medius* and *rectangularis*), which are sometimes crossed by small sandy nodular structures. Towards the top, at the transition with siliceous shales, *Dimorphograptus confertus* Lapworth and *D. confertus* cf. *swanstoni* Elles & Wood are found, showing a difference in thickness for the first part of the Silurian between the two localities. No trace of the black marker bed can be seen at Bou Leggou.

These sections give rise to a problem in the appreciation of the precise age for the base of the silts. However, given the usual conditions of sedimentation between the end of the Ordovician and the first Silurian and the fact that there is no proof of *A. acuminatus* at the beginning of its biozone, one can, for cartographical purposes, take the Silurian as beginning with the silts. It remains to analyze the mineralogy of the black marker beds, and perhaps also the siliceous shales, to see whether they reflect volcanic activity, even if only very distant from this district of the eastern Anti-Atlas.

(4) *On the northern slope of the central High Atlas, at Ghogoult (Locality 4) and east of Tiwghaza (Locality 4')*. The important Hercynian tectonics which are manifest in the central High Atlas, formerly known as the 'Mountains to the East of Marrakech' (Roch 1939) or 'Demnate Atlas' (Lévêque 1961), do not enable us to establish a sure succession for the boundary in this part of Morocco. The Silurian with *A. acuminatus* is present in the allochthonous inliers of Aït Mallah and Aït Mdioual (geological map Azilal 1:100 000, 1985) and in the autochtonous deposits to the west of Tiwghaza (boundary of topographical sheets Telouat and Skoura 1:100 000).

In Aït Mallah, *C. normalis*, *D. modestus*, *A. acuminatus*, *C. vesiculosus*, *Monograptus revolutus* s.l. (Kurck), *Pribylograptus incommodus* (Törnquist) and *A.* cf. *atavus* have been identified; in Aït Mdioual, only the lower third in argillaceous or argillaceous-siliceous shales, with a very thin cleavage (overlain by drier, resonant shales, sometimes with drifted micas), and higher coarser beds with *C. cyphus*. The relations with the Ordovician cannot be defined since the earlier Silurian 'constitutes a level of preferential disharmony' (Jenny & Le Marrec 1980).

West of Tiwghaza, *D. modestus*, *A. acuminatus* and *C. vesiculosus* are recognized from the base of the first 5 m of sandy, coarse, micaceous shales underlying siliceous and phthanitic ones of the Llandovery succession. Jenny & Le Marrec (1980) described the last three metres of the upper Ordovician as composed of classic 'massive or irregular decimetrical sandstones-quartzites, sometimes with oscillation-ripples, whitish colour with dark patina and black microbrechic or microconglomeratic sandstones or clays with round and matt quartz'.

(5) *In the substratum of the Plateau des Phosphates (Locality 5)*. An oil-boring—BJ 105—on the geological map Qasbat Tadla (1:100 000, 1985, at x = 417·7, y = 216·8) terminated at a depth of 1017 m in the upper Ashgill. In a fragment of core between 963 to 988·5 m, in an argillaceous, graphitic, more or less siliceous facies, the lowest associations contain: (a) more argillaceous than siliceous beds with many slip planes with *C. normalis*, *C. rectangularis*, *D. modestus*, *A. acuminatus*, followed by (b) a more siliceous layer with the same association underlying the *C. vesiculosus*, *Dimorphograptus* and *C. cyphus* Zones. Although information is insufficient to define the boundary, a sudden change in facies (here between 988·5 and 1017 m) is found, with the same pattern as in other areas.

(6) *In the Moroccan central massif, Azrou area, at Bou-Ourarh (Locality 6)* (Topographical Sheet Aïn Leuh, 1:50 000, at about x = 503·5, y = 302·5). The Silurian here occurs as a siliceous facies alternating with real phthanites weathering light grey. It is the 'Formation dite de Mokattam' of Choubert (1956). It always lies upon ridges of sandy or even quartzitic material, which are more resistant in the landscape, and which can be assigned to the upper Ordovician without more precision in dating. Graptolites are found more or less at the contact. At one locality, there is *C. normalis*, *C. medius*, *C. rectangularis*, *C. vesiculosus*, *A. acuminatus*, *Glyptograptus* sp.

or *Orthograptus* sp., *P. incommodus*, *A.* ex gr. *atavus* and *Raphidograptus toernquisti* (Elles & Wood). The beds with *A. acuminatus* are less compact than those with *C. vesiculosus*. Rhuddanian and Aeronian rocks with the *Coronograptus gregarius* Zone are found down a small valley. Sandy layers occur at several levels in the Mokattam Formation and the sequence is repetitive. It is now known that this area has suffered greatly through Hercynian tectonism, so it seems that Bou-Ourarh is constructed of a number of tectonic slices in which the Silurian has often played the role of soapstones, and so it is not possible to find any Silurian beds conformably against the Ordovician sandstones. Although this district is not important for the boundary definition, it is a supplementary paleogeographical marker for the distribution of the *A. acuminatus* Zone.

(7) *In the Palaeozoic inliers of the north middle Atlas.*

(a) Tazekka (Eastern Morocco) (Locality 7). The same tectonics as at Bou-Ourarh cause repetition of the upper Ordovician and lower Silurian. At Souk et Tleta des Zerarda (Topographical Sheet Ribat el Kheir, 1:50 000, at x = 594·5, y = 373·7) at the top of the usual quartzites, almost vertical upper Ashgill black argillaceous-siliceous and siliceous beds contain *C. normalis*, *C. medius*, *C. rectangularis*, *C.* probably *longifilis* Manck, *C.* probably *trifilis* Manck, *D. modestus* and *A. acuminatus*. Silurian beds follow, but not quite in the same section.

(b) Immouzer du Khandar (Locality 8). The same situation exists at the NW end of the Immouzer du Khandar inlier (Topographical Sheet Sefrou, 1:100 000, at about x = 540·1, y = 353·7), where *A. acuminatus*, *C. normalis*, *C. miserabilis*, *C. rectangularis* and *D. modestus* are found in the argillaceous facies of the Mokattam Formation, but the locality is altered and schistosed, with bedding plane thrusts. This has contact with sandy pelites and big well-rounded quartzites of the upper glacial Ordovician, which are equivalent to the Upper Deuxième Bani Formation (Upper Ashgill) of the Anti-Atlas.

Conclusions

The base of the Silurian is seen in many areas of Morocco, and invariably in argillaceous facies, underlying sandstone levels and never in true phthanites. It is remarkable that in these sections no Ordovician faunas have been found, except at Moulay bou Anane. However, in the central Anti-Atlas, Tagounite area, at Jbel Larjame and at Oued Moulili, some badly preserved brachiopods are known from the upper part of the Upper Deuxième Bani Formation; these are from a more western region (south flank of Jbel Addana, south of Akka) and consist of *Hirnantia sagittifera* (M'Coy), *Eostropheodonta squamosa* Havlíček, and *Plectothyrella chauveli* Havlíček, from grits above microconglomeratic clays. These faunas are very important in dating the intra-Hirnantian tillite, and in these areas of the central Anti-Atlas the Silurian begins with a hardground followed by graptolites of later Llandovery age. We do not expect to find more significant faunas in the Ordovician rocks and future studies must turn to the sedimentology of the glacial phenomena and the volcanic influences in the eastern Anti-Atlas; and also to the description and illustration of the graptolites themselves.

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The Ordovician–Silurian boundary in the Algerian Sahara

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Synopsis

Two sections, at eastern Tassili-n-Ajjer and at El Kseib, demonstrate the Ordovician–Silurian boundary, with graptolites at intervals and rare shells, however the *acuminatus* Zone itself is not recorded. The sections are internationally important firstly in demonstrating excellent glacial and periglacial sediments during the late Ashgill, and secondly in showing that this continental ice-mass melted and was succeeded by, but was not the origin of, the transgression during the latest Ordovician, in *persculptus* Zone times.

Introduction

Because of the uplift that probably affected most of the Algerian Sahara near the end of the Ordovician, and the circumpolar conditions which caused the development of a continental ice sheet (Debyser *et al.* 1965), the Algerian Sahara seemed originally an unlikely country for biostratigraphical study of the Ordovician–Silurian boundary. However, detailed observations from the boundary beds enable us to show clearly an almost continuous succession from the Ordovician to the Silurian in the eastern Tassili-n-Ajjer, whereas to the west, in the Ougarta range, there is a probable hiatus. Moreover, these observations suggest some interesting conclusions about the palaeogeography because this is a country where the glacial events are particularly striking (Fig. 1).

Eastern Tassili-n-Ajjer sections of the Djanet–In Djerane Oued tray and of the In Djerane Oued

Kilian (1928) drew attention to this area by pointing out the presence of a fauna of lowermost Llandovery age. Unhappily, this discovery was forgotten and it was many years later when interest was aroused again following a preliminary collection by the 'Mission sédimentologique sur la couverture sédimentaire du Boudin saharien' in 1965. Two further studies were carried out in the field (1978, 1982) despite substantial logistical difficulties; but only some of the successive results have been published, others are in press.

The stratigraphical succession is as follows (Fig. 2):
Above the Gara Tembi sandstones with a glacial relief:

(*a*) the Arrkine argillaceous sandy formation (about 90 m) in which a new fauna with *Climacograptus* (*Climacograptus*) *gelidus* nov. sp., *C.* (*Climacograptus*) *arrikini* nov. sp. and *C.* (*Climacograptus*) *normalis ajjeri* Legrand occurs near the base.

(b) The shaley formation of Oued In Djerane in which the following distinctions can be made:

Lower member (80 m) of silty claystones and siltstones with a few carbonate levels; the fauna is as follows: *C.* (*Climacograptus*) *normalis ajjeri* Legrand, *C.* (*Climacograptus*) *pseudovenustus* Legrand, *C.* (*Climacograptus*) *pretilokensis* nov. sp., *C.* (*Climacograptus*) *tilokensis* Legrand and *Zygospiraella* sp.

Middle member (about 110 m) with: a lower submember of shales with *C.* (*Climacograptus*) *normalis ajjeri* Legrand, *Diplograptus* (?) *kiliani* Legrand; an upper submember of siltstones and silty shales with *C.* (*Climacograptus*) *freuloni* nov. sp., *C.* (*Climacograptus*?) *incommodus* nov. sp., and *Glyptograptus* (*Glyptograptus*) *sahariensis* nov. sp. and near the top *C.* (*Climacograptus*) *imperfectus* Legrand, and ?*G.* (*Glyptograptus*) aff. *persculptus* (Salter).

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Fig.1

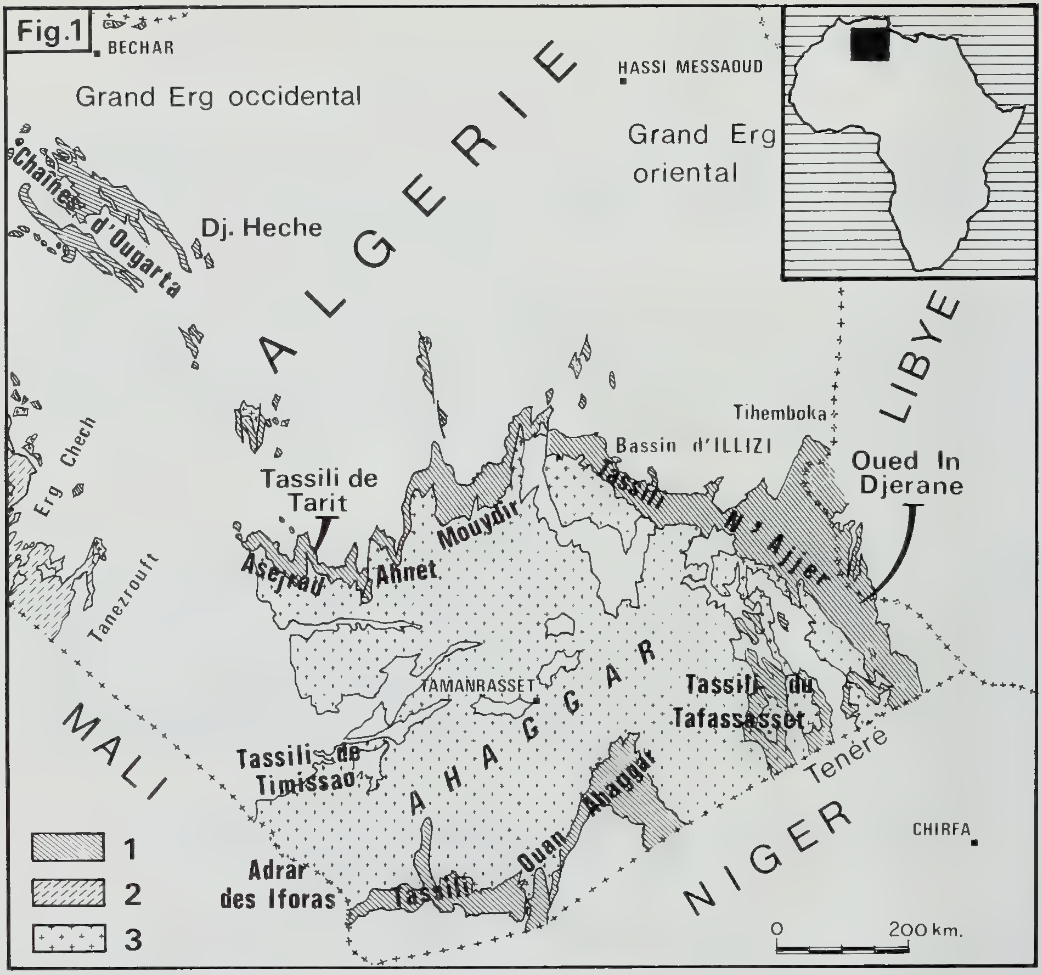
BECHAR
Grand Erg occidental
ALGERIE
HASSI MESSAOUD
Grand Erg oriental
Chaînes d'Ougarta
Dj. Heche
Erg Chech
Tassili de Tarit
Tanezrouft
Aseirad
Ahnet
Mouydir
Tihemboka
Bassin d'ILLIZI
Tassili N' Ajjer
Oued In Djerane
LIBYE
TAMANRASSET
AHAGGAR
Tassili du Tafassasset
Tassili de Timissao
Adrar des Iforas
Tassili Ouan Ahaggar
Tenere
MALI
NIGER
CHIRFA
0
200 km.
1
2
3

Fig. 1 Outcrops of lower Palaeozoic in Algeria apart from the Intermediate Series (1); Intermediate Series and Cambro-Ordovician of the syneclise of Taoudeni (2); and Precambrian and Intermediate Series (3).

Upper member of sandstones with argillaceous silty intercalations. Fossils are only found near the base and include *Diplograptus africanus* Legrand, and *G.* (*Glyptograptus*) *tariti* Legrand and then, above, *Diplograptus fezzanensis* Desio.

A lower Llandovery age was originally suggested for the whole Oued In Djerane Formation (Legrand 1976, 1981, 1985*a*); then an Ordovician–Silurian boundary level at the top of the *Diplograptus* (?) *kiliani* Zone was proposed (Legrand 1985*b*, 1986), but a further possibility, of a boundary at the top of the Middle member, must be considered. The arguments in favour of this last possibility are as follows:

(i) A new subspecies very near to *Diplograptus* (?) *kiliani* is known in the Kurama Range, Usbekistan (but not in Kazakhstan) and it occurs, according to T. N. Koren, not below the *Parakidograptus acuminatus* Zone, as formerly believed, but below some beds where *C.* (*Climacograptus*?) *extraordinarius* or *G.* (*Glyptograptus*) *persculptus* was collected.

(ii) On the other hand, *C.* (*Climacograptus*) *incommodus* has some affinities with *C.* (*Climacograptus*) *extraordinarius* and in this respect the position of *Zygospiraella*, a genus

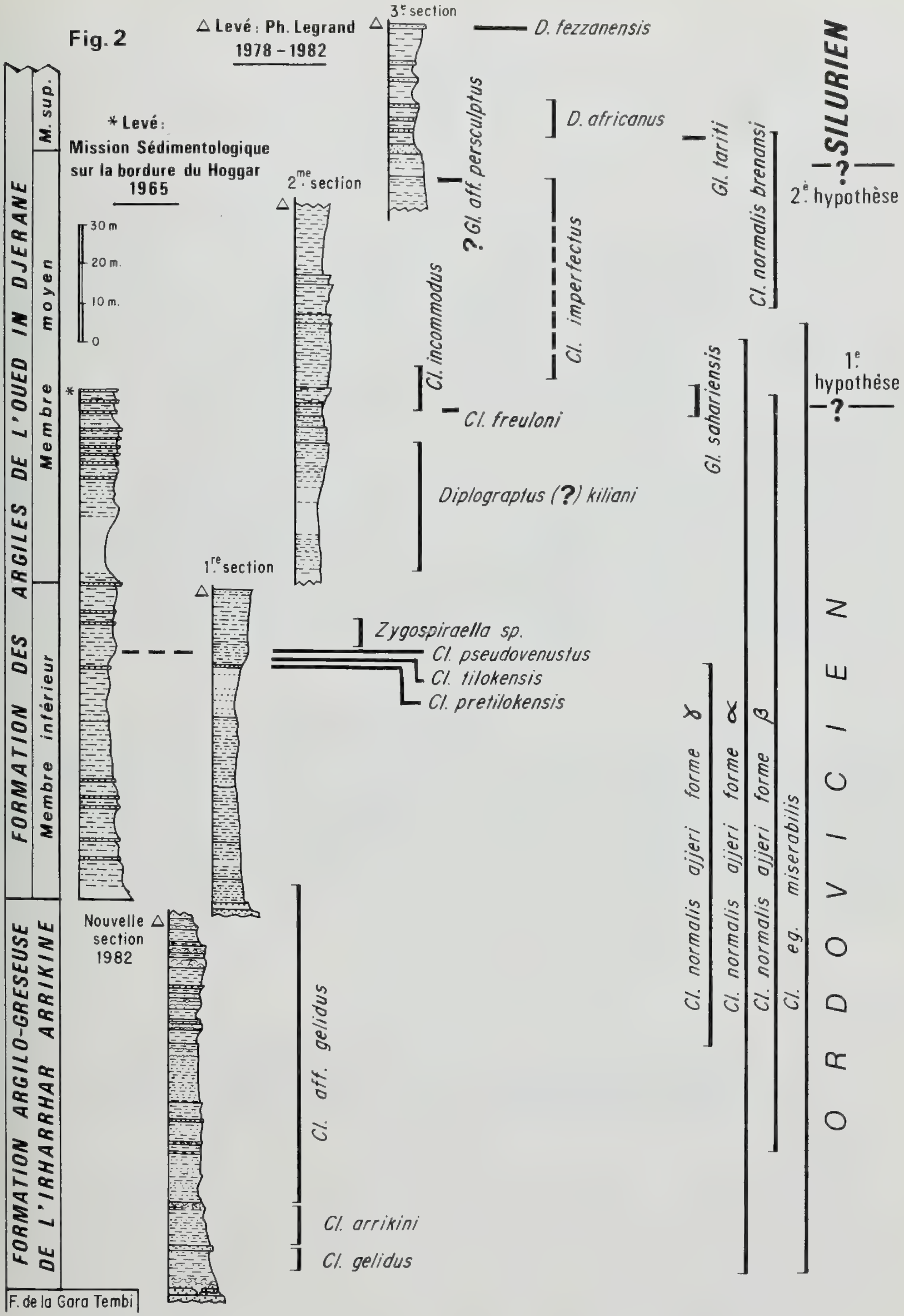
Fig. 2
△ Levé : Ph. Legrand 1978 – 1982
* Levé : Mission Sédimentologique sur la bordure du Hoggar 1965
3e section
2me section
1re section
Nouvelle section 1982
30 m
20 m.
10 m.
0
FORMATION DES ARGILES DE L'OUED IN DJERANE
M. sup.
Membre moyen
Membre inférieur
FORMATION ARGILO-GRESEUSE DE L'IRHARRHAR ARRIKINE
F. de la Gara Tembi
D. fezzanensis
?Gl. aff. persculptus
D. africanus
Cl. imperfectus
Cl. incommodus
Cl. freuloni
Diplograptus (?) kiliani
Zygospiraella sp.
Cl. pseudovenustus
Cl. tilokensis
Cl. pretilokensis
Cl. aff. gelidus
Cl. arrikini
Cl. gelidus
Gl. tariti
Cl. normalis brenansi
Gl. sahariensis
Cl. normalis ajjeri forme γ
Cl. normalis ajjeri forme α
Cl. normalis ajjeri forme β
Cl. eg. miserabilis
SILURIEN
?
2e hypothèse
1e hypothèse
?
ORDOVICIEN

Fig. 2 Distribution of the principal faunas in the sections of the Djanet–In Djerane Oued tray and the In Djerane Oued, Algeria.

only so far definitely recorded from the Silurian, would be the same as that in Kazakhstan (Oysu River section).

(iii) Finally, rare specimens of ?*G.* (*Glyptograptus*) aff. *persculptus* have been gathered just below the top of the middle member of the Oued In Djerane Formation.

The objections to the hypothesis are the following:

(i) *G.* (*Glyptograptus*) *sahariensis* is very close to *G.* (*Glyptograptus*) *tariti* and has the aspect of a Silurian *Glyptograptus.*

(ii) *Diplograptus africanus* seems to belong to the *Coronograptus cyphus* Zone (Legrand 1976), and consequently there is a very small thickness for the *Parakidograptus acuminatus* Zone and the *Cystograptus vesiculosus* Zone. The sandstones that form the top of the middle member may be thought to be the equivalent of the zone.

(iii) *Parakidograptus acuminatus* has not yet been found; one can think of the sandstones that form the top of the middle member as the equivalent of the biozone characterized by this species. However, nor has it been found near the Libyan boundary, where the shales take the place of the sandstones owing to the later transgression there, and where the sedimentation seems to have been more continuous.

(iv) Perhaps in this apparently very confined area the vertical range of species many not have been absolutely the same as in less restricted regions.

To conclude, two hypotheses can be proposed for the position of the Ordovician–Silurian boundary, but the highest seems the most likely. Moreover, there is no characteristic fauna of the Ordovician in the lower part of the section and this sets problems of correlation with the standard sections (Dob's Linn, Kolyma River, Yangtse Valley), and consequently this section in Algeria can only be a local reference. On the other hand, it has important palaeogeographical significance since it shows the beginning of the transgression onto the southeastern part of the Saharan shield before the end of the Ordovician, which must have involved the melting of the continental ice sheet, at least locally, before the beginning of the Silurian (Legrand 1985).

Ougarta Range–El Kseib section

In the Ougarta Range, the stratigraphical succession of the upper part of the Ordovician includes the argillaceous sandy Bou M'haoud Formation, which is overlain by the argillaceous sandy Jebel Serraf Formation. A mappable unconformity separates these two formations (Arbey 1962; Gomes Silva *et al.* 1963; BRP *et al.* 1964; Legrand 1974). In the eponymous locality, where that formation seems the most complete, the upper part of the Bou M'haoud

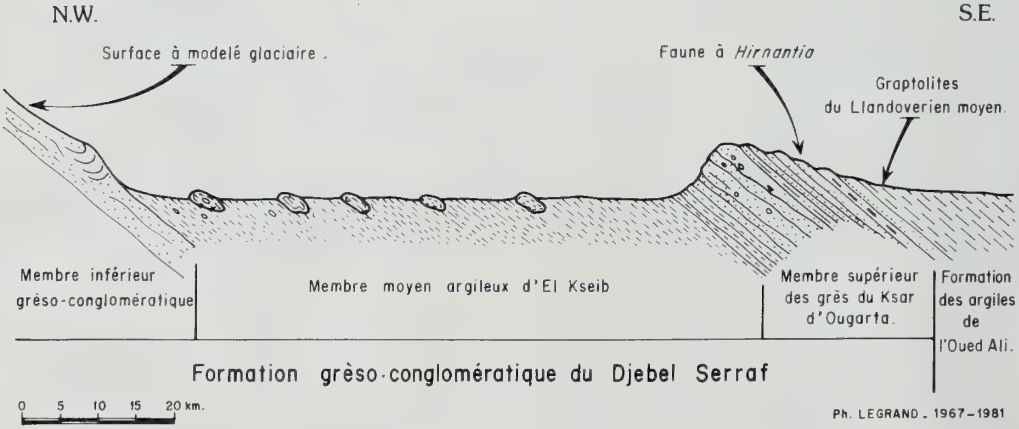
N.W.
S.E.
Surface à modelé glaciaire .
Faune à *Hirnantia*
Graptolites du Llandoverien moyen.
Membre inférieur grèso-conglomératique
Membre moyen argileux d'El Kseib
Membre supérieur des grès du Ksar d'Ougarta.
Formation des argiles de l'Oued Ali.
Formation grèso-conglomératique du Djebel Serraf
0 5 10 15 20 km.
Ph. LEGRAND . 1967–1981

Fig. 3 Section in the vicinity of the Ordovician–Silurian boundary at El Kseib, Ougarta range, Algeria.

Formation is apparently of Lower Caradoc age, with *Kloucekia* (*Kloucekia*?) nov. sp., *Calymenella* sp., *Drabovinella grandis* Mergl, and *Drabovia* sp.

At first this fauna was attributed to the Upper Caradoc and the beds from which it was collected were considered to belong to the lower member of the formation subjacent to the Jebel Serraf Formation. Going to the north west (in the Daoura), the succession is apparently complete up to the lower Ashgill. Above this the Jebel Serraf Formation appears to be absent or very thin in Bou M'haoud village, with siltstones and sandstones (channel deposits), but no fossils have been found. The quality of the outcrops does not allow us to see the contact with the lowest Silurian shales. Thus, it is near Ougarta that the Ordovician–Silurian boundary must be investigated.

In the classical El Kseib section discovered by Menchikoff (1930), the Bou M'haoud Formation is reduced to its lower member. Above, the Jebel Serraf Formation consists of a well-developed sandy, conglomeratic lower member, then the microconglomeratic shales of El Kseib that prove a periglacial environment; and above these, the sandstones of the 'Ksar d'Ougarta', It is at Ougarta that some brachiopods were gathered from this member by Poueyto (1950). Unhappily this fauna (which has been recollected since 1961) is poorly diversified and consists of *Plectotyrella chauveli* Havlíček, *Hirnantia* aff. *sagitiffera* (M'Coy), *Lingulella* sp., *Pseudobolus* sp., *Conchilolites* sp. and a homalonotid pygidium. The age of this member is uppermost Ashgill (Destombes 1971; Legrand 1974, 1985*a*, *b*). Above this the Oued Ali formation is found, whose base is characterized by a ferrugineous sandstone with ferrugineous nodules and then a bed of sandstone; there follows some varicoloured shales and coarse shaly sandstones with *C.* (*Climacograptus*) sp., and the member ends with black shales with *C.* (*Climacograptus*) aff. *rectangularis* M'Coy, *Orthograptus* aff. *mutabilis* Elles & Wood, ?*P.* (*Metaclimacograptus*) *phrygonius* Törnquist, and *Rastrites* sp., indicating a Middle Llandovery age.

Although this section is only interesting from a local point of view for the definition of the Ordovician–Silurian boundary, it has the wider advantage of showing that the glacial or periglacial environment ended just before the end of the Ashgill.

Conclusions

The Algerian Sahara is surprisingly important in increasing our knowledge of the Ordovician–Silurian boundary period. Studies in eastern Tassili-n-Ajjer show, in an almost continuous section through coastal sediments, the nature of the endemic faunal succession, which, however, has some affinities with southern Siberia. A palaeogeography can be drawn showing the area more or less neighbouring the South Pole, and the observations in the Ougarta Range strongly suggest the almost complete melting of the Upper Ordovician continental ice sheet before the Silurian transgression. This leads us to reconsider the importance of the melting in the mechanism of the transgression (Legrand 1985).

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The Ordovician–Silurian boundary in Mauritania

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Synopsis

Three sections are described across the Ordovician–Silurian boundary in Mauritania, each bearing well-developed glacial deposits succeeded by graptolitic shales. In general, fossils of the latest Ordovician and earliest Silurian are absent, apart from the southeastern section between Aratane and Oualata, at a cliff in Hodh, where the *persculptus* and *atavus* Zones are recorded.

Introduction

Three areas in Mauritania (Fig. 1) shed some light on the question of the Ordovician–Silurian boundary; however, the pioneer stage of work in these large areas encourages caution. The areas are:

1 Zemmour Noir (northern Mauritania), known from the masterly contribution of Sougy (1964) and included in the northern flank of the Reguibat uplift in Deynoux *et al.* (1985).

2 The Mauritanian Adrar, monographed by Trompette (1973), in the western part of the Taoudeni Basin (Deynoux *et al.* 1985).

3 Hodh, whose Precambrian and Ordovician glacial deposits were studied by Deynoux (1980); this is in the eastern extension of Tagant, which reaches the Adrar towards the S and SE. The Hodh escarpment frames a Cambro–Ordovician–Silurian ribbon to the N of the southern margin of the Taoudeni Basin before the post-Palaeozoic oversteps it (Deynoux *et al.* 1985).

In each area, the glacial upper Ordovician has been carefully studied and these deposits are more remarkable than those of Morocco, since they were nearer to the Lower Palaeozoic pole, and so record even more glacial activity, and, moreover, the glacial episode lasted for a longer time. The Ordovician–Silurian relationships are very gradual at Hodh and marked by an acute change of facies at Adrar and Zemmour.

Regional descriptions

1 **Zemmour Noir** (Fig. 2A, but chiefly Deynoux *et al.* 1985: 347, fig. 4; 354, fig. 6; and 369, fig. 7). The upper Ordovician consists of the Garat el Hamoueid Group and overlies rocks of Precambrian to Llanvirn age. Its upper boundary is correlated with the upper Ashgill by analogy with comparable deposits in Morocco and Algeria and its thickness varies between 0 and 200 m. The rocks are typical glacial deposits but these characteristics become less clear to the NW in the Dhlou Chain because of tectonic complications. Some sedimentological features suggest a more periglacial regime near the top. Faunas are very rare and consist only of 'indeterminable *Camarotoechia*' compared by Havlíček (1971) with other brachiopods of the upper sandstones of the Deuxième Bani of Morocco; and of *Cornulites.*

The base of the Silurian is marked by a very sharp discontinuity, and the system is well developed on the eastern margin of Zemmour, striking SSW–NNE. It always starts with *Demirastrites triangulatus* (Harkness) (determined by A. Philippot) in a facies of black, argillaceous, and some micaceous, shales. Its thickness seems to decrease evenly from 30 m in the north to 6 m in the south.

Among the detailed sections of Sougy (1964), the more northern, west of Gara Bouya Ali, has its base concealed by about 27 m of sandy 'oued': in the 3 m of overlying shales there are specimens of *Monograptus sedgwickii* (Portlock) (determined by A. Philippot), while a 30 cm bed of sandstones separates the top of the Garat el Hamoueid Group from the hidden part.

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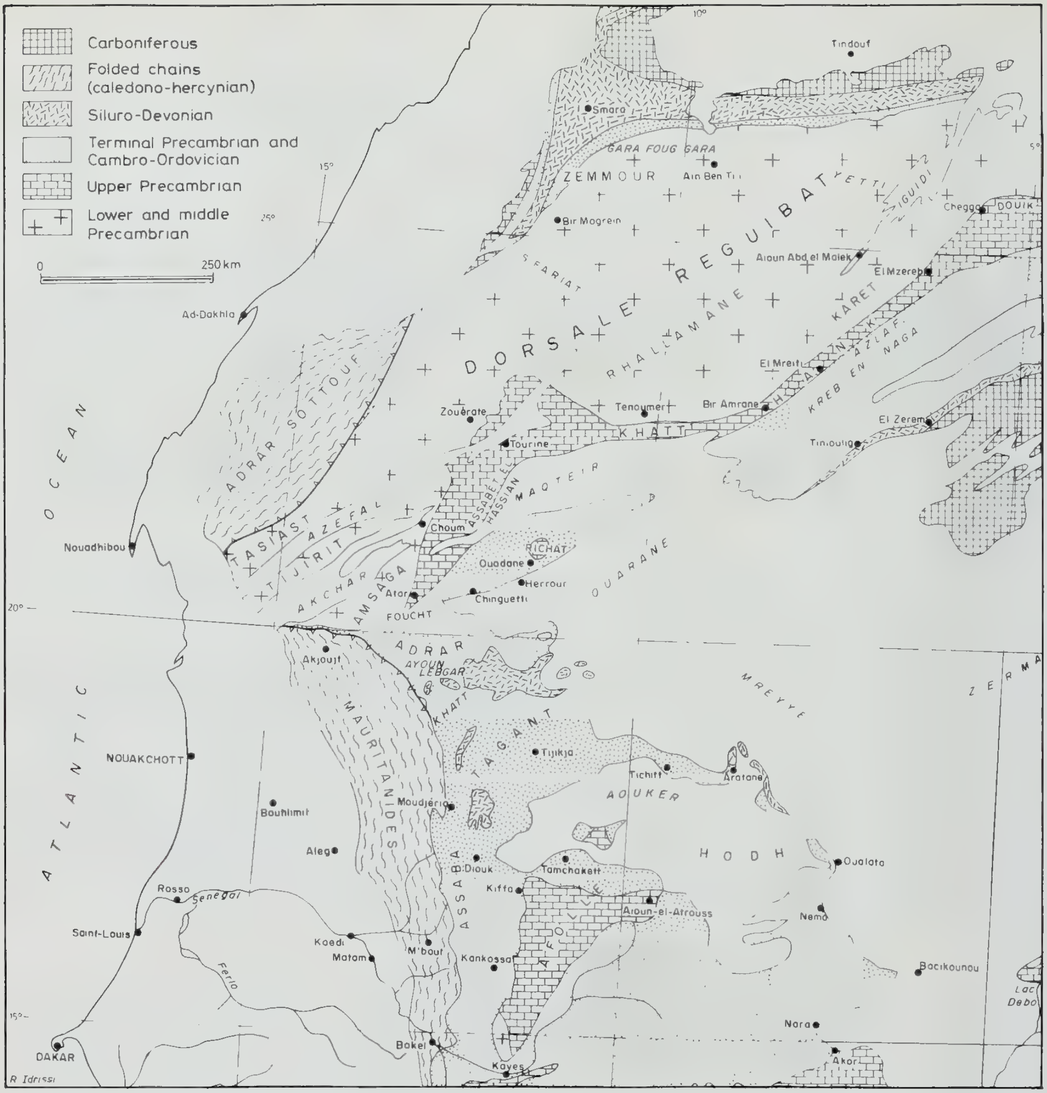
Carboniferous
Folded chains (caledono-hercynian)
Siluro-Devonian
Terminal Precambrian and Cambro-Ordovician
Upper Precambrian
Lower and middle Precambrian
0
250 km
Tindouf
Smara
GARA FOUG GARA
ZEMMOUR
Ain Ben Tili
Bir Mogrein
DORSALE REGUIBAT
YETTI
IGUIDI
Chegga
DOUIK
Aioun Abd el Malek
El Mzereb
SFARIAT
RHALLAMANE
KARET
HANK
AZLAF
KREB EN NAGA
El Mreiti
Bir Amrane
Tenoumer
KHATT
El Zerem
Tinioulig
Zouerate
Tourine
Ad-Dakhla
ADRAR SOTTOUF
OCEAN
ATLANTIC
TASIAST
AZEFAL
TIJIRIT
AKCHAR
AMSAGA
ASSABET EL HASSIAN
MAQTEIR
Choum
RICHAT
Ouadane
Herrour
OUARANE
Atar
Chinguetti
FOUCHT
ADRAR
AYOUN LEBGAR
KHATT
Nouadhibou
20°
15°
25°
10°
5°
Akjoujt
MAURITANIDES
TAGANT
Tijikja
Tichitt
Aratane
AOUKER
MREYYE
ZERMA
NOUAKCHOTT
Boutilimit
Moudjeria
Aleg
HODH
Oualata
Rosso
Senegal
Saint-Louis
Ferlo
ASSABA
Diouk
Tamchakett
Kiffa
Aioun-el-Atrouss
AFOLLE
Nema
Koedi
Matam
M'bout
Kankossa
Bacikounou
Lac Debo
Nara
Bokel
Kayes
Akor
DAKAR
R Idrissi

Fig. 1 Geological sketch of the western margin of the Taoudeni Basin, Mauritania, after Deynoux (1980).

Elsewhere, the surface of the sandstones at the contact with the shales is sometimes covered by a yellow coating. At Gara Foug Gara there is 2 m between '*Camarotoechia*' and *Demirastrites triangulatus*. There is therefore not much hope of defining the boundary exactly in Zemmour Noir, unless new discoveries are made in the western tectonized part. The Silurian has been noted in the Dhlou Chain but has not been systematically studied.

2 **The Mauritanian Adrar** (Fig. 2B, but chiefly Deynoux *et al.* 1985: 371, fig. 11; 374, fig. 12; 378, table 3). This area geomorphologically consists of (roughly from NNE to SSW), the Atar plain, the cliff, the plateaus (tabular zone) and the SW margin (folded zone), overlapped by the Mauritanides chain. The Ordovician–Silurian boundary is exposed in the two last units, but the area can be treated as a whole, whilst noting that the Silurian becomes more sandy to the

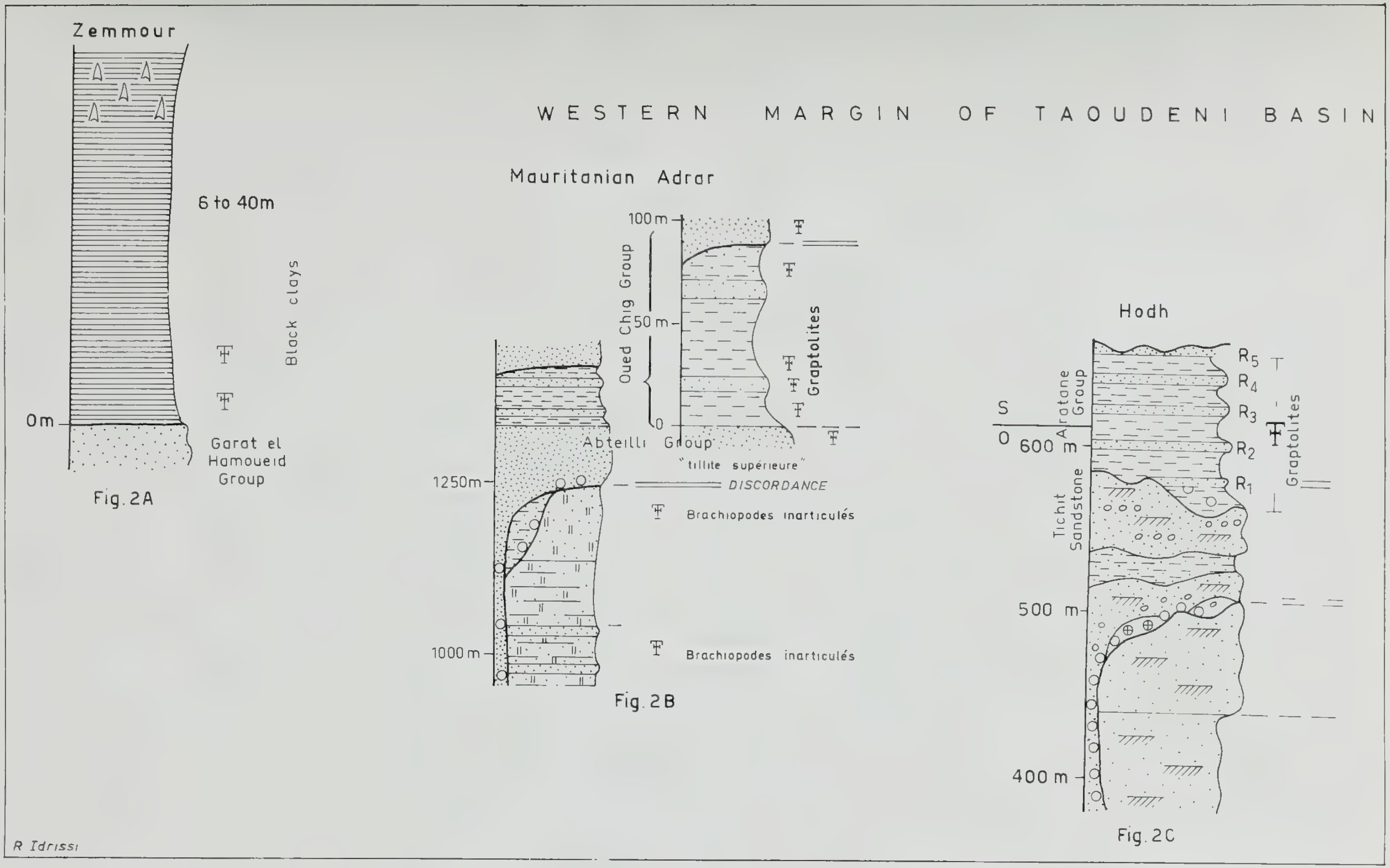
WESTERN MARGIN OF TAOUDENI BASIN
Zemmour
6 to 40m
Black clays
0m
Garat el Hamoueid Group
Fig. 2A
Mauritanian Adrar
100 m
50 m
0
Oued Chig Group
Graptolites
Abteilli Group
"tillite supérieure"
1250m
DISCORDANCE
Brachiopodes inarticulés
1000 m
Brachiopodes inarticulés
Fig. 2B
Hodh
S
O
600 m
Aratane Group
Tichit Sandstone
R5
R4
R3
R2
R1
Graptolites
500 m
400 m
Fig. 2C
R Idrissi

Fig. 2 Ordovician–Silurian boundary sections in Mauritania. The boundary is between R_2 and R_3 sandstones in the Hodh section (after Deynoux *et al.* 1985).

WSW. The glacial formation and the Silurian have been called 'Supergroup 3' by Trompette (1973), subdivided into the Abteilli Group and the Oued Chig Group.

(a) The Abteilli Group represents the glacial upper Ordovician whose lower boundary is difficult to establish because the glacial deposits occur in a landscape long exposed to continental deposition and weathering. The only earlier marine palaeontological horizon consists of lingulids of probable Cambro-Ordovician boundary age (determined by P. Legrand). The top of the group is marked by sandy eskers which reflect the withdrawal of the land ice to the south-east. At the time of Monod's survey (1952) in this district, some brachiopods in a sandstone from the folded zone at Ayoun Lebgar were determined by D. Le Maitre, who recognized the genera *Camarotoechia*, *Rhynchonella* (especially *R.* ex gr. *borealis*), *Orthis*, *Dalmanella* etc., but frequently with nomenclatural doubt. Monod thought that these sandstones were of Silurian age and that influenced the palaeontologist in her attribution to a high level in the Wenlock. However, these brachiopods may perhaps better be compared with those from Gara Foug Gara. J. Drot considers that in Zemmour as well as in Adrar all these fossils are indeterminable, but it is tempting to compare the total fauna directly. In the section, collected again by Trompette (1973), the usual graptolitic shales are immediately above the brachiopod-bearing lenticular sandstones, which indicate a marine incursion which might have been contemporaneous with those of Zemmour or the upper sandstones of the Deuxième Bani, and so Trompette has suggested that they belong to the lower Silurian. However, prudence is necessary with such weak data and both possibilities remain hypotheses.

(b) The base of the Oued Chig Group. In the fifteen sections and complementary support sections, Trompette (1973) was able to verify the concordance between the Abteilli Group and the Oued Chig Group and also the striking difference in sedimentation between the two groups. Their contact is rarely clear: there is often 1 m or more of sandy debris masking the extreme base of the Silurian. The oldest graptolites are: *Climacograptus normalis* Lapworth, *C.* cf. *rectangularis* (M'Coy), *C.* cf. *scalaris* (Hisinger), ?*C.* sp. or *Pseudoglyptograptus* sp., cf. *Pseudoclimacograptus* (*Metaclimacograptus*) *hughesi* (Nicholson), *Diplograptus magnus* Lapworth, *D. modestus* Lapworth or *D. magnus*, *Pristiograptus regularis* (Törnquist), *Lagarograptus tenuis* (Portlock), *M. sedgwickii* and ?*Cyclograptus* sp. or *Calyptograptus* sp. There is no *Akidograptus acuminatus* (Nicholson) but a part of the Rhuddanian may be present when the lowest association contains only the first *Climacograptus* and *Diplograptus* either *modestus* or *magnus*. In Adrar it appears that the Llandovery Series begins earlier than in Zemmour because of the scarcity of monograptids at the base.

3 **The Hodh** (Fig. 2c, but chiefly Deynoux *et al.* 1985: 389, fig. 16 and unpublished determinations). The subdivisions adopted here are Tichit Sandstones for the glacial formation and Aratane Group for the sandstones and shales with graptolites. The definition of the Ordovician–Silurian boundary (Cocks 1985) may modify somewhat the Silurian attribution of some of the basal graptolitic sediments.

The glacial complex rests on any formation among those defined as Cambro-Ordovician. The major erosional disconformity which opens the glacial cycle is perhaps also in places an angular unconformity, for example in Tagant (Dia *et al.* 1969). Deynoux (1980) has recognized a lower and an upper part in a total thickness of the order of 100–150 m. The upper part, with several members, includes sandstones and microconglomeratic clays underlying a landmark sandstone R_1, followed by sandy clays (still with microconglomeratic layers) under a second sandy landmark R_2, above which are the clays with graptolites of the Aratane Group. To the east there are further sandstones termed R_3 and R_4. This group ranges from 100–130 m in thickness.

In the more southeastern section, about halfway between Aratane and Oualata, a bed with graptolites between R_1 and R_2 contains some diplograptids identified as amplexograptids of Ashgill type. Following the escarpment to the north and west, the sandy landmarks become less easy to correlate but the zone of *Glyptograptus persculptus* is well represented:

(a) The more western layer, a portion of the Aratane cliff, appears to be deposited in a glacial gully under R_1 and contains only *Climacograptus normalis* and *C. transgrediens* Waern.

(b) The *persculptus* Zone contains: *Glyptograptus persculptus* (Salter), ?*Acanthograptus* sp. or ?*Koremagraptus* sp., *C. normalis*, *C. miserabilis* Elles & Wood, *C. transgrediens*, *C.* cf. *praemedius* Waern, *C. medius* (Törnquist), *C.* cf. *rectangularis*, *C.* cf. *indivisus* Davies, *C. minutus*? Elles & Wood, a more amplexograptid than climacograptid new form which recalls some figures of *Comatograptus* Obut & Sobolevskaya or *Hedrograptus* Obut, although more oval; rare fragments of *Orthograptus* ex gr. *truncatus* Lapworth, and ?*Akidograptus* sp. Some climacograptids show basal spines (Elles & Wood 1906; series of species of Manck 1924 (see Münch 1952); reminiscent of more ancient species such as those described by Ross & Berry, 1963). The septa of *G. persculptus* begins at the 4th theca.

These beds, except one, are in the portion of the Oualata-cliff, therefore to the NW–SE and above R_2 (but Deynoux cannot always decide between R_1 and R_2 towards the NW) in a facies of argillaceous shales and sandy layers and lenses, and some more micaceous beds.

(c) Above in the same member and in the portion of Oualata-cliff:

(i) A layer in a more sandy facies: *C. normalis*, *C. transgrediens*, *C. medius*, *C.* probably *praemedius*, the amplexograptid form, a proximal part of *Rhaphidograptus*?, a proximal part of *Akidograptus*? and some monograptid thecae.

(ii) In the same facies as (b): *C. normalis*, *C. miserabilis*, *C. minutus*, amplexograptid form narrower than those above, *Orthograptus truncatus abbreviatus* Elles & Wood, *Dimorphograptus* sp., *Pribylograptus incommodus* (Törnquist) and *Atavograptus* ex gr. *atavus* (Jones).

(iii) *C. normalis*, *C. miserabilis*, *Pseudoclimacograptus* (*Metaclimacograptus*) *hughesi* or *undulatus* (Kurck), *Diplograptus modestus*, *D. diminutus* Elles & Wood, and a single *Peiragraptus* or pathological specimen of *Diplograptus* sp.?

(d) The landmark bed R_3 is above these layers, except in one section where it has not been recognized (*C. normalis*, *P.* (*M.*) *hughesi*, *Dimorphograptus* cf. *confertus* Lapworth), and the same facies as (b) begins again with *C. normalis*, *C. rectangularis*, *P.* (*M.*) *hughesi* or *undulatus*, *D. modestus*, *Glyptograptus* ex gr. *tamariscus* (Nicholson), *G. tamariscus linearis*? Perner, *G.* either *angulatus* Packham or *distans* Packham, ?*Raphidograptus* sp., *A. atavus*, *A. strachani* Hutt & Rickards, *Lagarograptus acinaces*? (Törnquist), and *Coronograptus cyphus*? (Lapworth).

To the north of Aratane, beyond the post-Palaeozoic cover, towards Mejahouda and in the vicinity of Tinioulig, Sougy & Trompette (1976) have sampled the usual climacograptids, *D. modestus*, *Cystograptus vesiculosus* (Nicholson) and *A.* ex gr. *atavus*. All these graptolites are often irregularly flattened, preserved in iron oxides or with a fragile black pellicule. There is never an impression of fusellar tissue. Their deposit is rarely homogeneous along the rhabdosome. Some layers contain brachiopods and numbers of other organic fragments.

The Ordovician–Silurian boundary is therefore situated between the sandy landmarks R_2 and R_3 in the east of the Hodh. *G. persculptus* terminates the Ordovician, *A. acuminatus* is only suspected, and the remaining Rhuddanian is well represented. One should not forget that these collections are the first made systematically from this adverse environment, and reflect limited field-work, which was part of a large programme executed in a short time and with no possibility of immediate revision. The cliff at Hodh, in the Oualata area, if it were more accessible, would nevertheless be a first-rate place for a parastratotype, since it records the end of the African glacial phenomenon and has a good Ordovician–Silurian transition.

Recently, Legrand (1986) has described in detail (before the choice of the boundary) the lower Silurian at Oued in Djerane, Algeria, and has recognized new taxa. There is certainly some correlation between the Hoggar margin and the west of the Taoudeni Basin. However, before defining an 'African' fauna, it would be very useful to demonstrate with more certainty the effects of diagenesis on the preservation of graptolites, the more so because sections in proteic tissues have revealed the ability of the cortical layers to trap exogeneous particles. These extraneous particles could, of course, modify considerably any part of a rhabdosome.

Conclusions

From the Hodh to the Adrar, the post-glacial transgression would seem to have begun in the Ordovician and extended towards the west in the earliest Silurian, arriving later in the

Zemmour. The cliff to the north-west of Oualata is the best exposure of the local Ordovician–Silurian boundary, though it is still necessary to fully describe and figure the graptolites and complementary faunas from there.

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Ordovician–Silurian boundary in Victoria and New South Wales, Australia

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Synopsis

The late Ordovician and early Silurian is often represented by an unconformity or otherwise by beds bearing graptolites: no significant shelly faunas are known. In Darraweit Guim, Victoria, and in the Forbes–Parkes area of New South Wales, there may be beds spanning the Ordovician–Silurian boundary without a break, but nowhere have both the *persculptus* and *acuminatus* Zones been found in a single, structurally uncomplicated, succession.

Introduction

Ordovician and Silurian rocks crop out extensively in the Lachlan Fold Belt of southeastern Australia (Figs 1 and 3). A variety of facies is represented, from deep marine chert, black shale and turbidites, to shallow marine mudstone and sandstone. Carbonates and volcaniclastics occur, associated with island arc-type andesites in central New South Wales. The turbidite—black shale—chert association often contains rich and diverse graptolite assemblages and conodonts, but virtually no shelly fossils. Mixed graptolite–shelly fossil assemblages occur in some of the volcaniclastic deposits, but the shallow marine carbonates only contain shelly fossils.

Sections in central and eastern Victoria

No single section spanning the Ordovician–Silurian boundary has yet been located in Victoria, although there is reasonably convincing evidence of a complete but fault-disrupted succession at Darraweit Guim, near Melbourne (Fig. 1). Poor exposure and deep weathering, and the scarcity of fossils in the Silurian rocks, are the main difficulties in locating further sections. Another limiting factor is due to the effects of the Benambran Orogeny, a major accretionary event which took place at about the Ordovician–Silurian boundary and produced the Wagga Metamorphic Belt in eastern Victoria (Cooper & Grindley 1982). The orogeny is marked by a prominent facies change, from black shale with or without turbidites, to massive mudstone or quartzite. East of the metamorphic belt, the facies change follows a break in sedimentation, which in some places was accompanied by folding.

No such break in sedimentation occurs in the Melborne Trough in central Victoria, but here the lithological contrast produced by the Benambran Orogeny is such that the boundary interval became the preferred site for strike faulting during the Middle Devonian Tabberabberan orogeny, thus causing considerable complexity in the boundary sections.

Darraweit Guim

The only apparently complete succession spanning the Ordovician–Silurian boundary in Victoria occurs at Darraweit Guim, a hamlet 46 km NNW of Melbourne (Fig. 1). It is situated near the western margin of the Melbourne Trough, a basin in which there is record of continuous marine sedimentation from early Ordovician to late Early Devonian time (VandenBerg & Wilkinson, *in* Cooper & Grindley 1982). The boundary sequence recognized by VandenBerg *et al.* (1984) consists of three units, the Bolinda Shale, Darraweit Guim Mudstone and Deep Creek Siltstone (Fig. 2).

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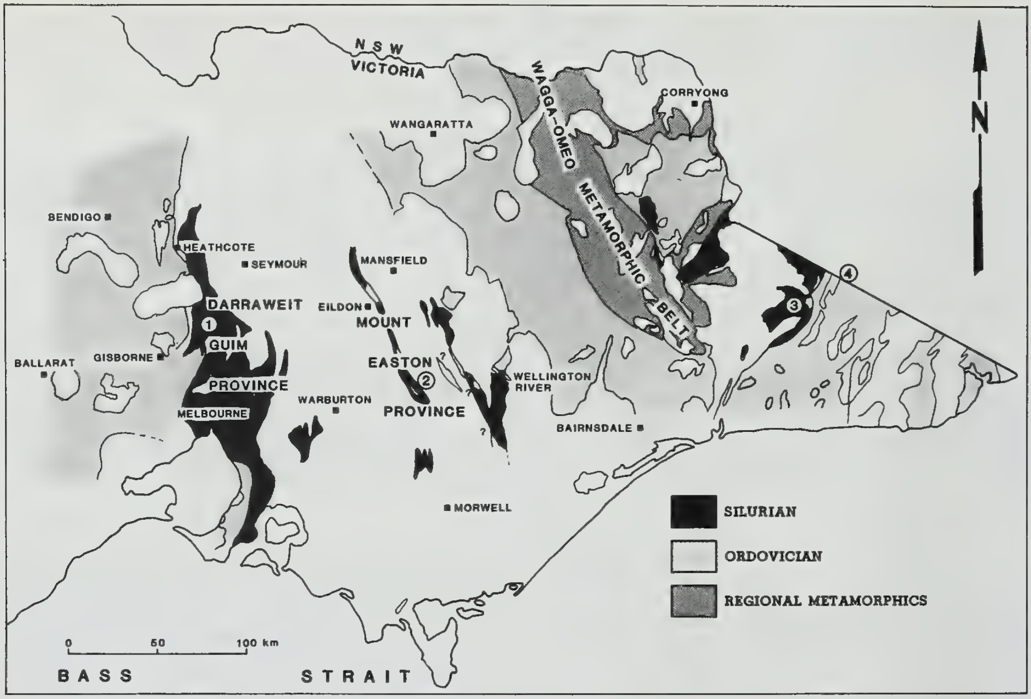
N S W
VICTORIA
WANGARATTA
CORRYONG
WAGGA-OMEO METAMORPHIC BELT
N
BENDIGO
HEATHCOTE
SEYMOUR
MANSFIELD
DARRAWEIT
1
GUIM
PROVINCE
EILDON
MOUNT
EASTON
2
PROVINCE
BALLARAT
GISBORNE
MELBOURNE
WARBURTON
WELLINGTON RIVER
BAIRNSDALE
3
4
MORWELL
SILURIAN
ORDOVICIAN
REGIONAL METAMORPHICS
0
50
100 km
BASS
STRAIT

Fig. 1 Distribution of Ordovician and Silurian rocks in central and eastern Victoria. Localities mentioned in text and Fig. 2 are: 1, Darraweit Guim; 2, Mount Easton region; 3, Yalmy River; 4, Delegate (southeast N.S.W.).

The Bolinda Shale is composed of 800 m or more of thin-bedded coarse-grained black shale and fine sandstone with a rich Bolindian graptolite fauna, comprising mostly cosmopolitan species. The assemblage consists of very abundant *Climacograptus latus*, *C. longispinus supernus* and *Orthograptus amplexicaulis* (sensu lato), somewhat less abundant *C. hastatus*, *C.* cf. *tubuliferus*, *Paraorthograptus pacificus pacificus* and *Dicellograptus ornatus*, and rare specimens of *Orthograptus fastigatus*, *Orthoretiograptus denticulatus* and *Pleurograptus linearis* (sensu lato). This assemblage constitutes the Zone of *D. ornatus* and *C. latus* of VandenBerg (*in* Webby *et al.* 1981) and is virtually identical to that of the *Paraorthograptus pacificus* Subzone at Dob's Linn (Williams 1982).

The overlying Darraweit Guim Mudstone consists of 20 to 45 m of sparsely fossiliferous black calcareous mudstone and slump-folded mudstone of partly evaporitic origin, and may be the only unit in Australia to show the effects of the late Ordovician glaciation (VandenBerg, in prep.). The impoverished shelly fauna consists of small bivalves, hyolithids, straight nautiloids, and a single trilobite, *Songxites darraweitensis.* More important, however, is the occurrence of *Climacograptus*? *extraordinarius* which is associated with *C. angustus* and *C.* cf. *acceptus* (VandenBerg *et al.* 1984). This assemblage represents the upper Bolindian Zone of *C.*? *extraordinarius* and is considered to correlate with the *C.*? *extraordinarius* Zone at Dob's Linn (Williams 1983).

Contacts between the Darraweit Guim Mudstone and the overlying Deep Creek Siltstone are usually poorly exposed and marked by bedding-parallel faults. The Deep Creek Siltstone is very thick (800–1000 m) and consists of poorly bedded, massive and bioturbated siltstone and thin rippled sandstone. Fossils are very rare. The lowest graptolite horizon occurs about 75 m above the base of the formation (and about 90 m above *C.*? *extraordinarius*) and contains *Glyptograptus* sp. (VandenBerg *et al.* 1984: fig. 11). A somewhat richer assemblage occurs 85 m

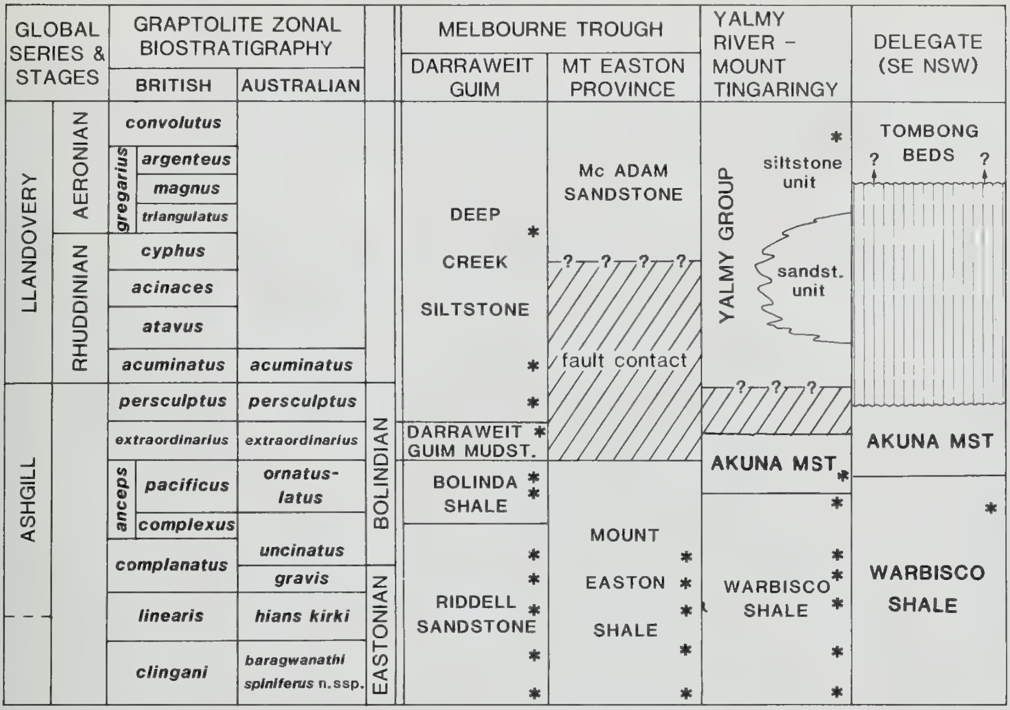
GLOBAL SERIES & STAGES
GRAPTOLITE ZONAL BIOSTRATIGRAPHY
BRITISH
AUSTRALIAN
MELBOURNE TROUGH
DARRAWEIT GUIM
MT EASTON PROVINCE
YALMY RIVER – MOUNT TINGARINGY
DELEGATE (SE NSW)
LLANDOVERY
AERONIAN
RHUDDINIAN
ASHGILL
BOLINDIAN
EASTONIAN
convolutus
gregarius
argenteus
magnus
triangulatus
cyphus
acinaces
atavus
acuminatus
acuminatus
persculptus
persculptus
extraordinarius
extraordinarius
anceps
pacificus
complexus
ornatus-latus
uncinatus
complanatus
gravis
linearis
hians kirki
clingani
baragwanathi
spiniferus n.ssp.
DEEP CREEK SILTSTONE
DARRAWEIT GUIM MUDST.
BOLINDA SHALE
RIDDELL SANDSTONE
Mc ADAM SANDSTONE
fault contact
MOUNT EASTON SHALE
YALMY GROUP
siltstone unit
sandst. unit
AKUNA MST
WARBISCO SHALE
TOMBONG BEDS
AKUNA MST
WARBISCO SHALE

Fig. 2 Correlation chart of Ordovician–Silurian boundary sections in Victoria. For location of columns, see Fig. 1. Graptolite horizons are shown by asterisks.

and 95 m higher in the same section (VandenBerg *et al.* 1984: fig. 3), and contains *Climacograptus normalis*, *C. angustus*, and *Glyptograptus*? *persculptus* or a species very close to it. This assemblage is considered to correlate with the British *G.*? *persculptus* Zone at Dob's Linn (Williams 1983).

The next graptolite zone, the Zone of *Parakidograptus acuminatus*, is based on a single described specimen of *P. acuminatus* cf. *acuminatus* (VandenBerg *et al.* 1984) which came from the core of an anticline north of Darraweit Guim, low in the Deep Creek Siltstone, but unfortunately structurally isolated from the more complete sections west of Darraweit Guim. Its precise stratigraphical relationship with the *G.*? *persculptus* Zone is therefore not known. The same applies to an assemblage from PL665, low in the Deep Creek Siltstone NW of Darraweit Guim, consisting entirely of *Glyptograptus*? *venustus* (Legrand *non* Mu) (figured as *C. normalis* in VandenBerg *et al.* 1984: fig. 10A).

Little work has been done on the sparse graptolite fauna higher in the Deep Creek Siltstone (Harris & Thomas 1937, 1949), and much of it is in need of revision. Sufficient material has been collected, however, to indicate that the graptolite record is far from complete and can only be correlated with reference to the standard British sequence.

Mount Easton

In the Mount Easton Province, farther east in the Melborne Trough (Fig. 1), VandenBerg (*in* Webby *et al.* 1981) has recognized a nearly complete Upper Ordovician sequence of graptolite faunas in the Mount Easton Shale (Fig. 2). Faunas range from the Darriwilian Zone of *Pseudoclimacograptus*? *decoratus* to the Bolindian Zone of *Dicellograptus ornatus* and *Climacograptus latus*. VandenBerg (1975) has recorded a possibly conformable relationship with overlying siltstone near Eildon, but elsewhere the shale is in fault contact with the 500 m thick McAdam

Sandstone (VandenBerg 1975). The latter contains a small late Llandovery graptolite assemblage including *Retiolites geinitzianus* (recorded as *Stomatograptus australis*), *Monograptus exiguus*, *M. turriculatus*, *M. spiralis permensus*, *M. priodon* and *M. pandus* (Keble & Harris 1934; Harris & Thomas 1947). There is a single record of Silurian graptolites, listed as *Glyptograptus tamariscus*, *Climacograptus* sp. and *Monograptus* spp. (Harris & Thomas 1954) from an outcrop adjacent to Mount Easton Shale in the structurally complex Mount Wellington Belt.

Eastern Victoria and the borderland with New South Wales

In the Yalmy River–Mount Tingaringy district in eastern Victoria (Fig. 1), the Warbisco Shale comprises about 500 m of black shale. This contains a graptolite sequence which is recorded by VandenBerg (1981) as complete from the Gisbornian Zone of *Nemagraptus gracilis*, to the Bolindian *D. ornatus*–*C. latus* Zone (Fig. 2). Locally, the black shale is overlain by a thin unit of sandstone and siltstone, the Akuna Mudstone, still with a full *D. ornatus*–*C. latus* zonal assemblage comprising *Dicellograptus ornatus*, *Climacograptus latus*, *C. longispinus supernus*, *C. hastatus*, *Paraorthograptus pacificus* and *Orthoretiograptus denticulatus*. This unit was formerly placed in the Yalmy Group (VandenBerg, *in* Webby *et al.* 1981: 33) but its relationship is not completely clear. In most places, the contact between Warbisco Shale and undoubted Yalmy Group is faulted, and the entire Akuna Mudstone is absent.

The 3700 m thick Yalmy Group consists of about 2700 m of siltstone containing very large lenses of deltaic? sandstone, overlain by about 1000 m of orthoquartzite turbidites (Fig. 2). Several small graptolite assemblages occur high in the siltstone unit, but only one has been studied sufficiently to permit correlation and it comprises *Petalograptus* sp., *Glyptograptus* sp., *Retiolites* cf. *perlatus*, and a variety of monograptids including *M. convolutus* which correlate with the mid-Llandovery *M. convolutus* Zone of Britain.

At Delegate in southeastern New South Wales, to the northeast of the Yalmy River–Mount Tingaringy district (Fig. 1), the 200–300 m thick Akuna Mudstone (R. A. Glen, in prep.) overlies the entire Warbisco Shale (Fig. 2). Most of the latter formation consists of black shale, ranging in age from Gisbornian (with *Climacograptus bicornis bicornis*) to Bolindian (with *C. latus* and *Orthograptus fastigatus*). A prominent facies change from black shale to grey-green siltstone occurs at the boundary with the Akuna Mudstone and may correlate with the transition from Warbisco Shale to Akuna Mudstone farther west. No fossils have been collected from the upper part of the Akuna Mudstone, but there is a good possibility that the unit extends into the Silurian.

The contact between the Akuna Mudstone and the overlying Tombong Beds is a low-angle unconformity, attributable to the Benambran Orogeny which, elsewhere in the same district, marks a period of strong folding (Glen & VandenBerg 1985, 1987). The Tombong Beds are thick and unfossiliferous, but a small graptolite assemblage has been recorded from the overlying Meriangaah Siltstone by Crook *et al.* (1973). They suggest a broad late Llandovery–early Wenlock age, based on the occurrence of *Retiolites geinitzianus angustidens*, '*Monograptus* cf. *auduncus*' (presumably *Monoclimacis adunca*), and *M.* ex gr. *priodon*.

Sections in central New South Wales

Similarly, in New South Wales no section has yet been demonstrated to exhibit a complete record of beds across the Ordovician–Silurian boundary. The main limiting factors are the poor exposure, the structural complexity and the lack of continuity of richly fossiliferous successions. Even in the tableland areas the topography is generally subdued, and the sequences are often deeply weathered. The effects of the latest Ordovician–early Silurian Benambran Orogeny are noticeable in many areas of New South Wales, as in eastern Victoria. This major event resulted in the closing of the Wagga Marginal Sea, and then of its deformation, metamorphism and plutonism to produce the upraised Wagga Metamorphic Belt (Fig. 3). No proven Silurian deposits are known to occur to the west of the Wagga Metamorphic Belt, and many areas to the east appear to have a less than complete record of deposition through the Ordovician–

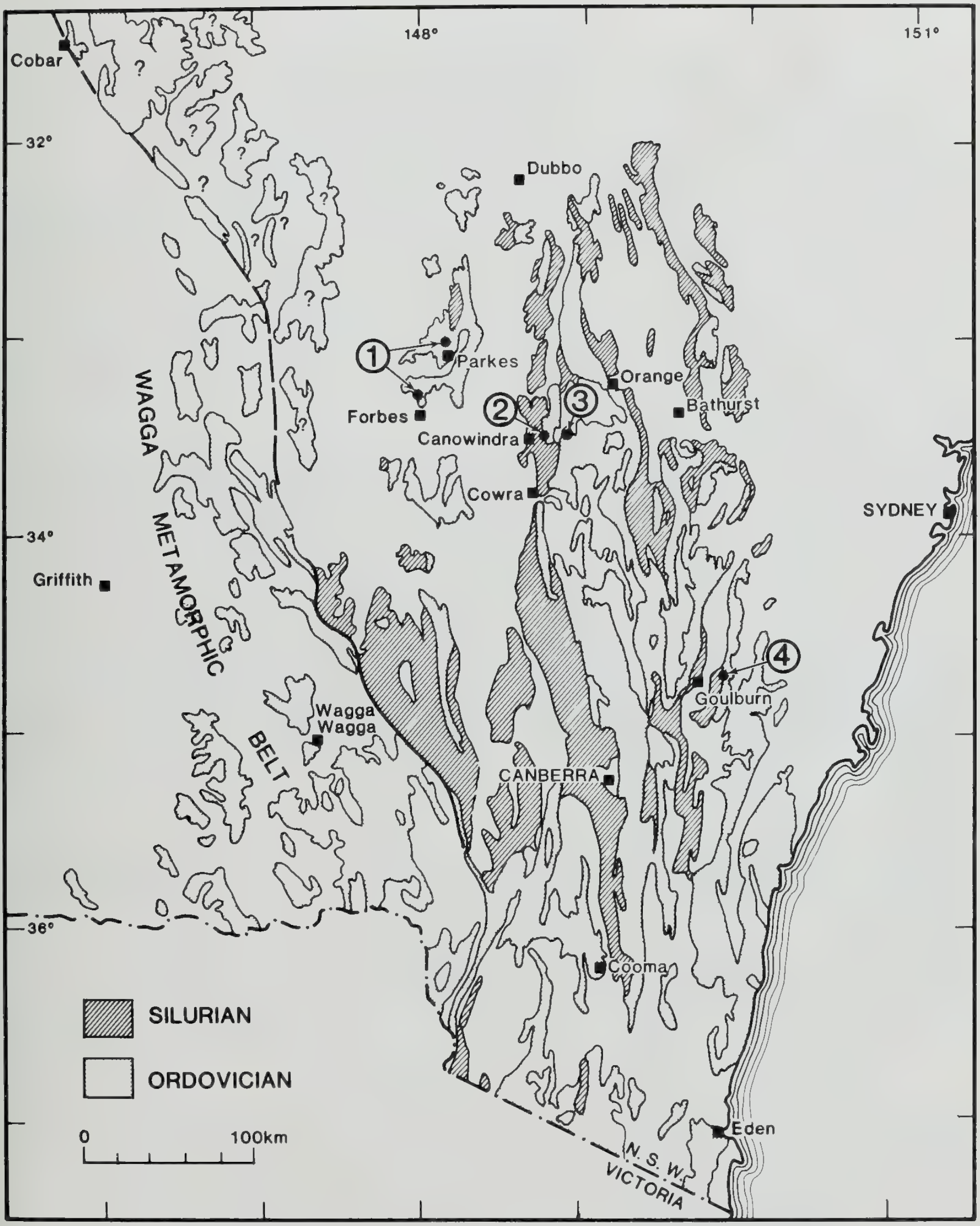
148°
151°
Cobar
32°
34°
36°
?
Dubbo
①
Parkes
Forbes
②
③
Orange
Bathurst
Canowindra
Cowra
SYDNEY
Griffith
WAGGA METAMORPHIC BELT
Wagga Wagga
④
Goulburn
CANBERRA
Cooma
Eden
N. S. W.
VICTORIA
SILURIAN
ORDOVICIAN
0
100km

Fig. 3 Map showing the distribution of Ordovician and Silurian rocks in central and southern New South Wales, and the location of Ordovician–Silurian boundary sections represented in Fig. 4.

Silurian boundary interval. The latest Ordovician deposits east of the Wagga Metamorphic Belt accumulated with associated graptolites in deeper waters as did much of the overlying Early Silurian, but many sections show physical breaks (unconformities, disconformities with associated facies changes or faults) reflecting the Benambran orogenesis or subsequent events.

The few sections which appear to show conformity unfortunately have an incomplete record of Late Ordovician to Early Silurian graptolite assemblages—late Bolindian occurrences followed by a significant barren interval to the succeeding mid-Llandovery assemblages, making it impossible to position the boundary closely (Figs 3–4). In addition to the rarity of proven early Llandovery deposits, there is an even greater paucity of established late Bolindian to early Llandovery shelly faunas. Indeed the graptolites are the only group to be adequately represented in the New South Wales successions. The sections with the best potential for establishing the Ordovician–Silurian boundary in New South Wales are in the Forbes area and east of Canowindra. Two less important sections occur in the Angullong–Four Mile Creek area and east of Goulburn.

1. *Forbes–Parkes.* The Cotton Siltstone of the Forbes area comprises separate exposures of a lower unit of late Ordovician age and an upper unit of Early Silurian age (Sherwin 1970, 1973) with an extensive strip of ground in between, representing unexposed intervening beds. Sherwin identified two graptolite assemblages from the lower unit, fauna A characterized by *Climacograptus supernus*, *C. hastatus*, *C. latus*, *Dicellograptus* cf. *elegans* and *Orthograptus truncatus* subsp., and assigned a Bolindian age; and fauna B typified by *C. normalis* and placed by Sherwin at or just above the Ordovician–Silurian boundary. The upper unit contains faunas C and D which are correlated with the late Llandovery (*sedgwickii* and *turriculatus* Zones); see also Sherwin (1974). *C. normalis* is the only determinable graptolite in fauna B and is a long-ranging species, and consequently can be of little use in establishing the position of the

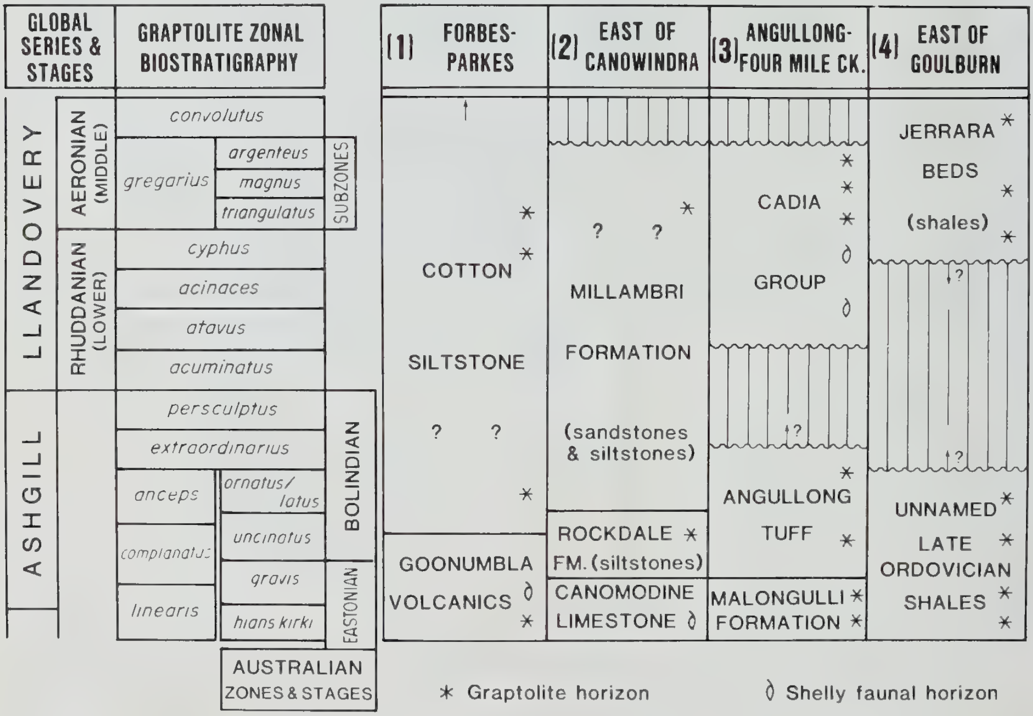
GLOBAL SERIES & STAGES
GRAPTOLITE ZONAL BIOSTRATIGRAPHY
LLANDOVERY
AERONIAN (MIDDLE)
RHUDDANIAN (LOWER)
ASHGILL
convolutus
gregarius
argenteus
magnus
triangulatus
SUBZONES
cyphus
acinaces
atavus
acuminatus
persculptus
extraordinarius
anceps
ornatus/latus
complanatus
uncinatus
gravis
linearis
hians kirki
BOLINDIAN
EASTONIAN
AUSTRALIAN ZONES & STAGES
(1) FORBES-PARKES
(2) EAST OF CANOWINDRA
(3) ANGULLONG-FOUR MILE CK.
(4) EAST OF GOULBURN
COTTON SILTSTONE
? ?
GOONUMBLA VOLCANICS
? ?
MILLAMBRI FORMATION
(sandstones & siltstones)
ROCKDALE FM. (siltstones)
CANOMODINE LIMESTONE
CADIA GROUP
?
ANGULLONG TUFF
MALONGULLI FORMATION
JERRARA BEDS (shales)
?
?
UNNAMED LATE ORDOVICIAN SHALES
✱ Graptolite horizon
∂ Shelly faunal horizon

Fig. 4 Correlation chart of Ordovician–Silurian boundary sections in central New South Wales.

boundary. Sherwin (*in* Pickett 1982) estimated the Cotton Siltstone of the Forbes area to be a total of 1500 m thick, and a large part of this is unexposed. For instance, only 100 m of the upper unit is well exposed in the road cutting and quarry near Cotton Trig north-west of Forbes (Sherwin 1973: fig. 4).

At 'The Secrets' north of Parkes, a 90 m thick sequence of the Cotton Siltstone includes several graptolite assemblages (Sherwin 1976) which do not occur near Forbes. These probably come from stratigraphical levels equivalent to the unexposed gap (between faunas B and C) of the Forbes section. The assemblages range in age from late lower to early middle Llandovery (*M. cyphus* to *M. triangulatus* Zones). The earliest assemblages, represented through the interval from 60–70 m on Sherwin's (1976: fig. 3) measured column, include elements such as *Climacograptus normalis*, *Pseudoclimacograptus* sp., *Glyptograptus* sp. and *Monograptus*? *strachani*. Unfortunately, however, there is as yet no evidence in the sections of the Cotton Siltstone near Forbes and Parkes of the presence of either the latest Ordovician graptolite zones of *C.*? *extraordinarius* and *G. persculptus*, or the earliest Llandovery zones of *P. acuminatus* or *C. vesiculosus*. Attempts are to be made to arrange the drilling of the unexposed part of the Forbes section, as it promises to provide the most complete, well preserved and structurally most uncomplicated record of graptolite assemblages through the Ordovician–Silurian boundary interval in Australia.

2. *East of Canowindra*. It is also possible that the Millambri Formation, as redefined by Ryall (1965), contains a continuous sequence of beds across the Ordovician–Silurian boundary but this 1240 m thick siliciclastic (poorly bedded arenite and well bedded siltstone) succession needs to be studied in much more detail. In its type area, in the core of the Cranky Rock Anticline east of Canowindra, Ryall (1965) has recognized the Millambri Formation as resting conformably on the Rockdale Formation. This siltstone unit has a Late Ordovician graptolite assemblage identified by Ryall (1965) as *Climacograptus bicornis* (probably erroneously), *C.* sp., *Dicellograptus* sp. and *Glyptograptus* sp. Judging from its stratigraphical relationships with the underlying Canomodine Limestone, the Rockdale Formation is unlikely to be older than early Bolindian (Webby *et al.* 1981). In a separate faulted sliver at Lidcombe Pools, to the east of the type area, the top of the Millambri Formation has produced a graptolite fauna of middle Llandovery age, that is about the level of the *M. gregarius* Zone. Elements of this fauna have been recorded by Percival (1976) as including *Glyptograptus tamariscus*, *Monograptus jonesi*, *Pseudoclimacograptus* (*Metaclimacograptus*) *hughesi*, *P.* (*M.*) *andulatus* and *P.* (*Clinoclimacograptus*) *retroversus*.

3. *Angullong–Four Mile Creek*. In the Angullong–Four Mile Creek area, Jenkins (1978) has found a late Bolindian assemblage in the uppermost part of the Angullong Tuff and referred the fauna of *Climacograptus supernus*, *C. latus*, *C. normalis* and *Dicellograptus ornatus ornatus* to the *D. anceps* Zone. Jenkins (1978) has also noted that the horizon lies beneath the top of the Angullong Tuff, so that volcanic activity may have continued somewhat beyond the end of *anceps* Zone time. These tuffs are succeeded disconformably by clastics and limestones of the Cadia Group, the basal part being judged by Jenkins to be about the level of the *C. vesiculosus* Zone. This implies a break of possibly two graptolite zones of the latest Ordovician and one of the earliest Silurian.

4. *East of Goulburn*. Sherwin (*in* Pickett 1982) has noted that while the Early Silurian shales of the Jerrara Beds east of Goulburn 'are closely associated with a great thickness of Late Ordovician strata of similar rock kinds, and because of structural uncertainties and known faults in this belt it is not known if sedimentation was continuous from Late Ordovician to Silurian times or not'. Graptolite assemblages of Bolindian and middle–late Llandovery ages have been recorded from many localities, and in one road section on the Hume Highway, a tightly folded succession of shales exhibits both Bolindian assemblages and Llandovery assemblages ranging from the *M. cyphus* to *M. convolutus* Zones (Creaser 1973). However, again there appears to be a significant break (or barren interval) representing the latest Ordovician (two zones) and the earliest Silurian (two zones).

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The base of the Silurian System in Tasmania

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Synopsis

The base of the Silurian System in Tasmania lies within the Westfield Sandstone, probably just below an horizon exposed in the road cutting immediately east of Westfield Quarry and containing a rich fauna including ?*Akidograptus*, *Atavograptus*, *Climacograptus normalis* and *Glyptograptus persculptus*.

Introduction

The base of the Silurian System in Tasmania lies within the uppermost formation of the Gordon Group, the Westfield Sandstone (this includes the Westfield Beds of Corbett & Banks 1974 and equals the Arndell Sandstone of Baillie 1979). The Gordon Group is a predominantly shallow water sequence, deposition of which began in the Canadian and continued apparently without interruption into the early Silurian. Within this group in the Florentine Valley (lat. 42° 37′ S, long. 146° 22′ E) the uppermost carbonate formation, the Benjamin Limestone, is overlain by the Westfield Sandstone. Stratigraphically equivalent limestones are overlain by siltstones and/or sandstones in the Linda Valley in western Tasmania and Mole Creek in northern Tasmania, but only in the Florentine Valley are the sequences sufficiently exposed, structurally simple enough and known well enough for consideration in the context of this volume.

The relevant sections in the Florentine Valley lie within the Westfield Syncline and the Tiger Syncline of the Florentine Synclinorium (Corbett & Banks 1974). These structures in the relevant areas appear to be simple and most of the dips lie between 30° and 50° (Fig. 1). The two areas of particular importance are the Westfield Syncline and the eastern flank of the Tiger Syncline.

Biostratigraphy

In the Westfield Syncline the top of the Benjamin Limestone, e.g. at Corbett & Banks (1974) locality 13, contains stromatoporoids (Webby & Banks 1976), rugose corals including *Foerstephyllum* sp., *Palaeophyllum* spp., *Favistina* sp., *Cyathophylloides* sp., favositids including *Palaeofavosites* sp., auloporids including *Eofletcheria* sp., heliolitids including *Calapoecia* sp. and *Coccoseris*, halysitids including *Catenipora* sp. and *Falsicatenipora* cf. *chillagoensis* (Etheridge), ?*Beloitoceras* sp., *Dinorthis* sp. (Laurie 1982) and the conodonts *Belodina compressa* and *Phragmodus undatus* (Banks & Burrett 1980). The assemblage suggests correlation with the *P. linearis* Zone (Webby *et al.* 1981) and is clearly Ordovician.

No contact between the Benjamin Limestone and the Westfield Sandstone is exposed. Localities F1 of Baillie & Clarke (1976) and C.&B.15 of Corbett & Banks (1974) are clearly close to the base of the Sandstone. F1 and F9 of Baillie & Clarke (1976) are closely similar faunally (see Table 1) as are GB15 and GB16 of Corbett & Banks (1974), and differences between F1 and F9 on the one hand and C.&B.15 and 16 on the other may be ecological rather than temporal since F1 and F9 are in sandstone and the other two in siltstone. The fauna from F3 of Baillie & Clarke (1976) is similar to that of C.&B.15 and 16 and is also in siltstone. All five localities can conveniently be grouped together as different from other and higher horizons. *Glossograptus* sp. and a trinucleid related to *Guandacolithus* suggest that these horizons are late Ordovician. A few metres stratigraphically above F1 is an horizon, L6 of Laurie (1982), containing *Hirnantia* sp. and *Isorthis* (*Ovalella*) n. sp. (Laurie 1982). A further 40 m stratigraphically higher is a richly fossiliferous horizon (C.&B.18, B.&C.F2, L11) with *Onniella* sp., *Eospirifer* sp., and other brachiopods, *Pterinea* sp., *Orthodesma* sp., *Encrinuraspis* sp., *Brongniartella* sp., *Eokosovopeltis* sp.,

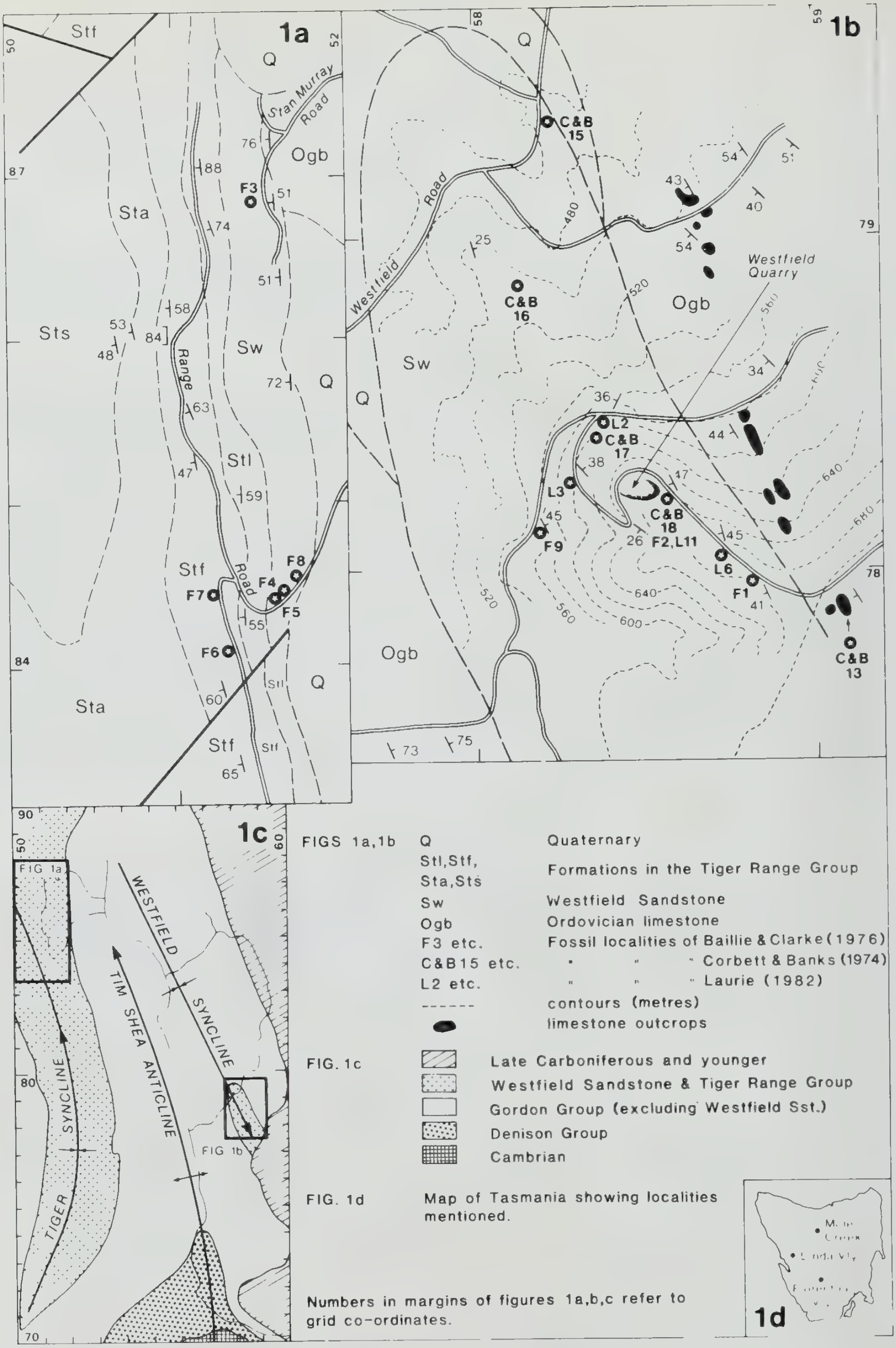
1a
1b
1c
1d
Q
Stf
Stl
Sta
Sts
Sw
Ogb
Stan Murray Road
Range Road
Westfield Road
Westfield Quarry
F3
F4
F5
F6
F7
F8
F9
F1
F2,L11
C&B 13
C&B 15
C&B 16
C&B 17
C&B 18
L2
L3
L6
WESTFIELD SYNCLINE
TIM SHEA ANTICLINE
TIGER SYNCLINE
FIG 1a
FIG 1b
FIGS 1a,1b
Q Quaternary
Stl, Stf, Sta, Sts Formations in the Tiger Range Group
Sw Westfield Sandstone
Ogb Ordovician limestone
F3 etc. Fossil localities of Baillie & Clarke (1976)
C&B 15 etc. " " " Corbett & Banks (1974)
L2 etc. " " " Laurie (1982)
------ contours (metres)
limestone outcrops
FIG. 1c
Late Carboniferous and younger
Westfield Sandstone & Tiger Range Group
Gordon Group (excluding Westfield Sst.)
Denison Group
Cambrian
FIG. 1d
Map of Tasmania showing localities mentioned.
Numbers in margins of figures 1a,b,c refer to grid co-ordinates.

Fig. 1 Ordovician–Silurian Boundary outcrops in Tasmania. 1a, The Tiger Syncline; 1b, The Westfield Syncline; 1c, The Florentine Valley, also showing the positions of Figs 1a and 1b; 1d, The Florentine and Linda Valleys and Mole Creek within Tasmania.

Table 1 Biostratigraphical range chart of fossils from the Westfield Sandstone, Tasmania.

Taxon	CB15	F3	F9	F1	L2	L3	L6	F4	CB16	CB18 F2 L11	F8	F5
Lepidocyclus	×	—	×	×	—	—	—	—	—	?	—	—
***Pterinea* sp. A P.&G.-T.	×	—	—	—	—	—	—	—	—	—	—	—
Onniella	×	—	×	×	—	—	—	×	—	×	×	—
*?*Onniella* n. sp. L.	—	—	—	—	—	—	—	—	—	×	—	—
cf. *Calymene birmanicus*	×	—	—	—	—	—	—	—	×	?	—	—
cf. *Guandacolithus*	×	×	—	—	—	—	—	—	×	?	—	—
cf. *Heterorthis*	—	×	—	—	—	—	—	—	—	—	—	—
Byssoconchia	—	×	—	—	—	—	—	—	—	—	—	—
Bumastus	—	×	—	—	—	—	—	—	×	—	—	—
Flexicalymene	—	×	—	—	—	—	—	—	—	—	—	—
?*Dalmanophyllum*	—	—	×	×	—	—	—	—	—	—	—	—
?*Holophragma*	—	—	×	×	—	—	—	—	—	—	—	—
Dolerorthis	—	—	×	×	—	—	—	—	—	—	—	—
Kjerulfina	—	—	×	—	—	—	—	—	—	×	—	—
**Hirnantia* n. sp. L.	—	—	—	—	×	×	×	—	—	—	—	—
**Isorthis* (*Ovalella*) n. sp. L.	—	—	—	—	—	—	×	—	—	—	—	—
**Kinnella* cf. *kielanae* T.S.	—	—	—	—	—	×	—	—	—	—	—	—
Bekkeromena	—	—	—	—	—	—	—	×	—	×	×	—
Hedstroemina	—	—	—	—	—	—	—	×	—	×	×	—
Orthodesma	—	—	—	—	—	—	—	—	×	×	—	—
Pterinea	—	—	—	—	—	—	—	—	×	×	—	—
***Tasmanoconularia* sp. Parfrey	—	—	—	—	—	—	—	—	×	—	—	—
Glossograptus	—	—	—	—	—	—	—	—	×	—	—	—
retiolitid	—	—	—	—	—	—	—	—	×	—	—	—
favositids	—	—	—	—	—	—	—	—	—	×	—	—
***Eospirifer* sp. S.&B.	—	—	—	—	—	—	—	—	—	×	—	—
Brongniartella	—	—	—	—	—	—	—	—	—	×	—	—
Bumastoides	—	—	—	—	—	—	—	—	—	×	—	—
Encrinuraspis	—	—	—	—	—	—	—	—	—	×	—	—
Encrinurus	—	—	—	—	—	—	—	—	—	×	—	—
Eokosovopeltis	—	—	—	—	—	—	—	—	—	×	—	—
Gravicalymene	—	—	—	—	—	—	—	—	—	×	—	—
**?*Akidograptus* B.B.&R.	—	—	—	—	—	—	—	—	—	×	—	—
***Atavograptus* B.B.&R.	—	—	—	—	—	—	—	—	—	×	—	—
***Climacograptus normalis* Lapworth	—	—	—	—	—	—	—	—	—	×	—	—
***Glyptograptus persculptus*	—	—	—	—	—	—	—	—	—	×	—	—
***Glyptograptus* cf. *persculptus*	—	—	—	—	—	—	—	—	—	×	—	—
***Eospirifer tasmaniensis* S.&B.	—	—	—	—	—	—	—	—	—	—	—	×

**Indicates published description and/or figure.
*Indicates figured and described in a Ph.D. thesis (Laurie 1982).
Other taxa names based on preliminary to somewhat detailed examination.
Records from Baillie (1979); Baillie, Banks & Rickards (1978); Baillie & Clarke (1976); Banks & Burrett (1980); Corbett & Banks (1974); Laurie (1982); Parfrey (1982); Pojeta & Gilbert-Tomlinson (1977); Sheehan & Baillie (1981); Webby & Banks (1976).

Bumastoides sp., *Gravicalymene* sp., ?*Akidograptus* sp., *Atavograptus* sp., *Climacograptus normalis* Lapworth, *Glyptograptus persculptus* (Salter) and *G.* cf. *persculptus*. The graptolites suggest either the *persculptus* Zone or an horizon low in the *acuminatus* Zone (Baillie *et al.* 1978). In view of the recent decision to place the base of the Silurian System at the base of the *acuminatus* Zone (Cocks 1985), this horizon must lie close to the base of the System.

Horizons (L2, L3 of Laurie) contain *Hirnantia* sp. and one of these also contains *Kinnella* cf. *kielanae* (Laurie 1982). The stratigraphical positions of these horizons are not clear and one or both could be stratigraphically below F2 (both are some tens of metres topographically lower).

The brachiopods *Bekkeromena* sp., *Hedstroemina* sp. and *Onniella* sp. have been collected from an horizon (F4 of Baillie & Clarke 1976) on the eastern flank of the Tiger Syncline. A slightly higher horizon (F5 of Baillie & Clarke) on the flank of the Tiger Range contains *Eospirifer tasmaniensis* Sheehan & Baillie (1981) in abundance. This occurs 65 m below the top of the Westfield Sandstone which is overlain by the Gell Quartzite and then the Richea Siltstone of the Tiger Range Group (Baillie 1979). The Richea Siltstone contains graptolites in an horizon 300 m above that with *E. tasmaniensis* and the graptolites indicate a very late Llandovery age (Baillie 1979).

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Stratigraphy and Palaeontology of the Ordovician–Silurian boundary interval, Anticosti Island, Quebec, Canada

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Synopsis

Anticosti Island provided the principal alternative boundary stratotype to Dob's Linn, Scotland, for the base of the Silurian System. It represents the best exposed, most fossiliferous, continuous section across the systemic boundary and has virtually all the attributes required of a stratotype. The 1100 m Upper Ordovician–Lower Silurian (Richmondian to Jumpersian stages) sequence of limestone with minor shale represents deposition in a marginal carbonate basin. The latest Ordovician Ellis Bay and earliest Silurian lower Becscie formations contain a record of eustatic sea level change and profound faunal changes. The seven members in the Ellis Bay Formation appear to reflect eustatic changes associated with the Saharan glaciation. The Ellis Bay–lower Becscie interval has yielded some 300 species of most invertebrate phyla. Correlation of this interval is best achieved through conodonts, ostracodes and palynomorphs, together with brachiopods and trilobites. There is a profound faunal change in conodonts and palynomorphs at 90 cm above the base of member 7, Ellis Bay Formation which is taken as the systemic boundary. Precise correlation of this level to the *P. acuminatus* graptolite Zone is difficult, but it probably lies at or just below this zonal level, somewhere within the upper *G. persculptus* Zone. The Anticosti sequence represents a standard reference for carbonate platform successions across the boundary and it also holds much information in regard to the processes and timing of the various faunal/floral extinctions which together form a Phanerozoic extinction event second in significance only to the terminal Permian event.

Introduction

The best exposed, most fossiliferous and complete section through the Ordovician–Silurian boundary interval occurs on Anticosti Island, Quebec. In these qualities as well as the lack of deformation, excellent preservation and diversity of faunas, Anticosti is comparable to other outstanding stratigraphical sections of Ordovician and Silurian strata such as the type Cincinnatian Series, the type Wenlock Series, the Silurian of Gotland and the type Pridoli Series. Dob's Linn and Anticosti–Gaspé were the only boundary sections formally visited by the Ordovician–Silurian Boundary Working Group, in 1979 and 1981 respectively. Arguments supporting Anticosti as a boundary stratotype were advanced by Barnes *et al.* (1981), Barnes & McCracken (1981*a*, *b*) and McCracken & Barnes (1981). The I.U.G.S., however, has ratified the decision of the Ordovician–Silurian Boundary Working Group to choose Dob's Linn, Scotland, as the boundary stratotype (Cocks 1985) and this issue is considered elsewhere in this volume. However, it is the view of this author, and others, that a serious error of judgement has been made in this decision and that reconsideration should occur in the near future (Lespérance *et al.* 1987). In this paper, a general review is presented of the stratigraphy and palaeontology of the boundary interval on Anticosti. Many data were presented by workers in the volumes prepared for the Anticosti field excursion edited by Lespérance (1981). Some additional data have been published in the intervening period and some new conodont data are presented herein.

Anticosti Island lies in the Gulf of St Lawrence and is approximately 200 km long and up to 50 km wide (Fig. 1). The only town is Port Menier on the western end which can be reached by plane (Québecair) from Sept Iles on the north shore, or by ferry from Rimouski on the south shore of the Gulf. The island has a network of logging roads, reflecting the main economic activity of the past fifty years. In 1975, the island was expropriated by the Province of Quebec and converted to a hunting and fishing reserve: it has over 70 000 deer and some of North

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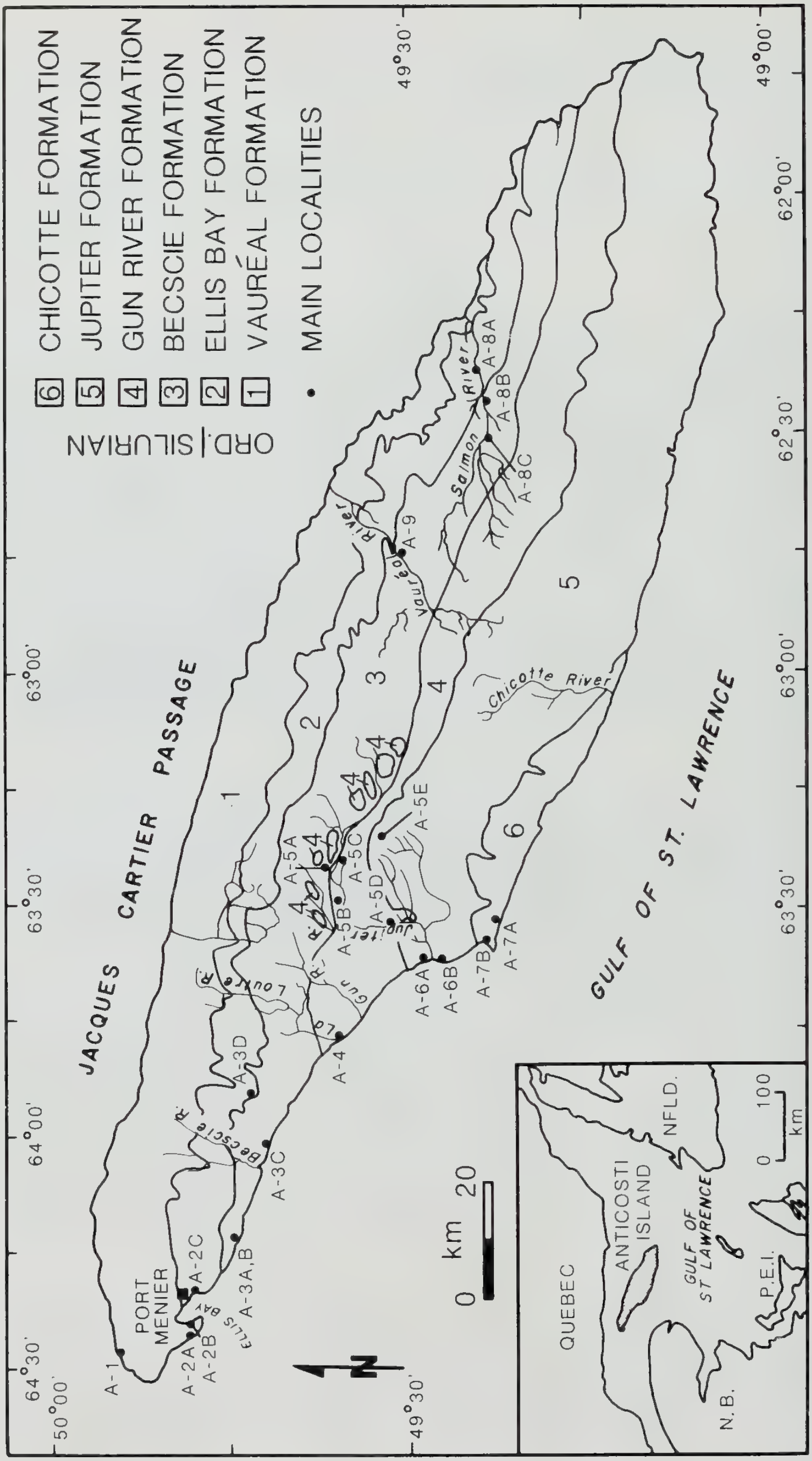
6 CHICOTTE FORMATION
5 JUPITER FORMATION
4 GUN RIVER FORMATION
3 BECSCIE FORMATION
2 ELLIS BAY FORMATION
1 VAURÉAL FORMATION
• MAIN LOCALITIES
SILURIAN
ORD.
JACQUES CARTIER PASSAGE
GULF OF ST. LAWRENCE
PORT MENIER
ELLIS BAY
Becscie R.
Loutre R.
Gun R.
Jupiter R.
Vauréal River
Salmon River
Chicotte River
A-1
A-2A
A-2B
A-2C
A-3A,B
A-3C
A-3D
A-4
A-5A
A-5B
A-5C
A-5D
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Fig. 1 Map of Anticosti Island showing distribution of formations and location of key sections described in detail in Barnes *et al.* (1981).

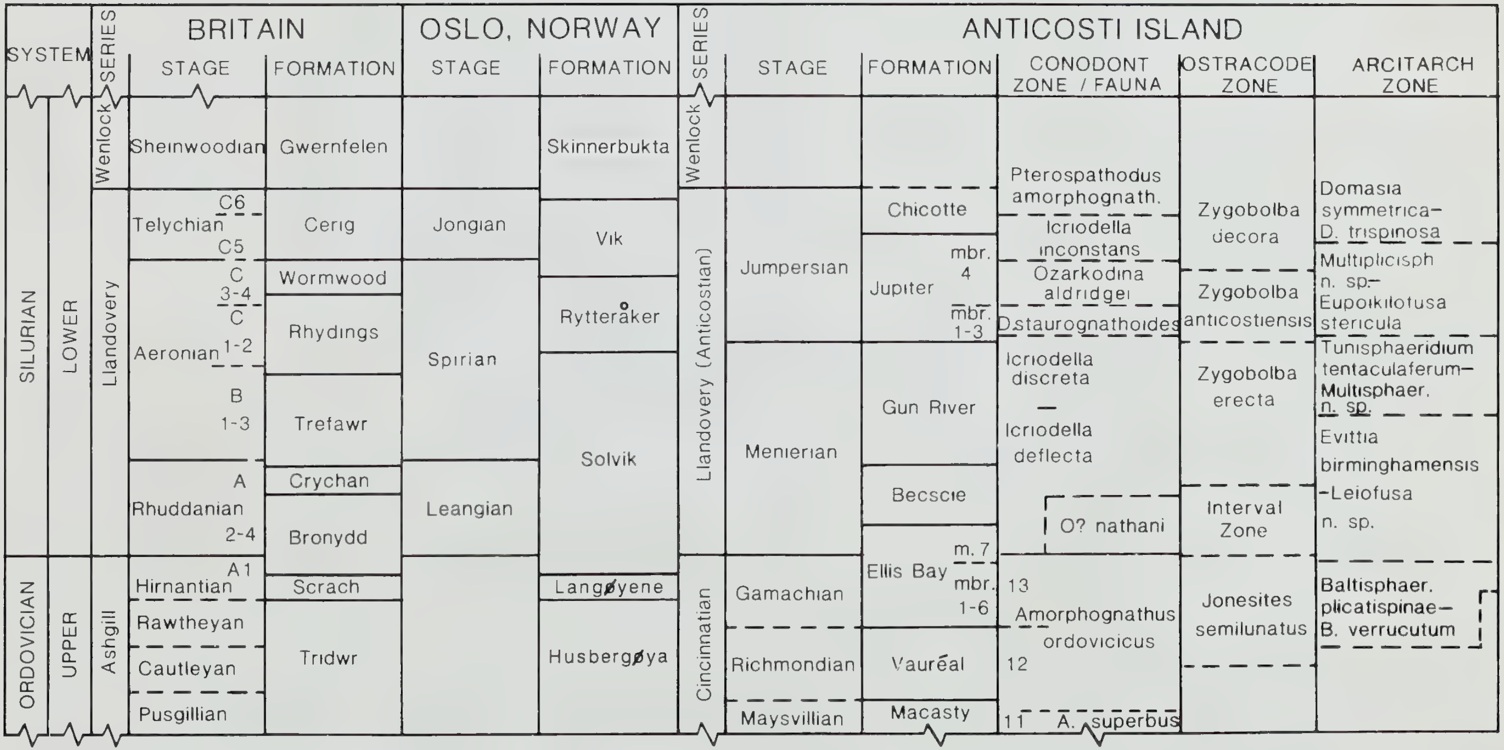
America's best salmon rivers. Port Menier has a hotel; cabins and camping facilities have been developed; vehicles may be rented, or ferried from Rimouski; travel to the eastern and central parts of the island requires a permit.

Stratigraphy

The island exposes an Upper Ordovician–Lower Silurian (Richmondian, Gamachian, Menierian, Jumpersian stages) succession, approximately 1100 m thick, comprising the Vauréal, Ellis Bay, Becscie, Gun River, Jupiter and Chicotte formations (Figs 1, 2). These limestones and minor shales and sandstones were deposited in the Anticosti Basin. Older parts of the succession are exposed as a discontinuous, narrow belt on the north shore of the Gulf, and in western Newfoundland. Offshore basinal equivalent strata are exposed to the south of the Logan's Line structural front in the Gaspé Peninsula. Oil exploration wells on Anticosti and seismic work south of the island have provided additional information on the regional stratigraphy (Roliff 1968; Petryk 1981*d*; Roksandic & Granger 1981). The strata dip at less than two degrees to the southwest and conodont colour alteration indices (CAI 1) indicate that burial temperatures did not exceed 80°C (Nowlan & Barnes 1987). Excellent exposure is present as cliff sections around the coast and as a wide wave-cut platform in some places; inland, exposures occur mainly along rivers and roadcuts. Key boundary sections are described by Barnes *et al.* (1981) and McCracken & Barnes (1981*a*); section numbers referred to in the former paper are also shown in Fig. 1. Space limitations do not permit a full review of previous studies (see references in McCracken & Barnes 1981*a*; Lespérance 1981). Key contributions on stratigraphy include those of Schuchert & Twenhofel (1910), Twenhofel (1914, 1928), Bolton (1961, 1972), Copeland & Bolton (1975) and Petryk (1979, 1981*a*).

During the Early and Middle Ordovician, the Anticosti Basin acted as a stable platform receiving shallow water carbonates. In response to tectonic activity of the Taconic Orogeny, the area was converted into a foreland basin first receiving the black shales of the Macasty Formation (Maysvillian), followed by 1100 m of shale and limestone of the Vauréal Formation. Only the upper third of the Vauréal outcrops at the surface on Anticosti and it forms most of the northern and western coastal outcrops. Bolton (1972) recognized that the units referred to the English Head and Vauréal Formations by Twenhofel (1921, 1928) belonged to the same formation; he proposed a lower shale and an upper limestone member. Petryk (1981*a*, *c*) recognized five informal members in the Vauréal Formation. Bolton's upper member, 150 m thick, consists of thin- to medium-bedded, grey, lime mudstone to skeletal wackestone with rare skeletal packstone, and interbedded grey shale. Intraformational limestone conglomerate and ball and pillow slump structures are common. Trace fossils are abundant; small coral-stromatoporoid bioherms occur near the top; some beds have concentrations of the stromatoporoid *Aulacera* (*Beatricea*) up to 3 m in length. Sedimentological data (Petryk 1981*a*) and conodont palaeoecology (Nowlan & Barnes 1981) indicate a general upward shallowing sequence. The numerous minor cycles in the relative abundance of the conodont genera *Drepanoistodus* and *Panderodus* (Nowlan & Barnes 1981: fig. 4) may represent climatic Milankovitch cycles which produced repetitive oceanic water mass interactions. The faunas of the Vauréal Formation suggest a Richmondian age (Fig. 2); the main study by Twenhofel (1928) was followed by others on graptolites (Riva 1969; Riva & Petryk 1981), ostracodes (Copeland 1970), chitinozoans (Achab 1977*a*, *b*), and conodonts (Nowlan & Barnes 1981).

The upper Vauréal and Ellis Bay Formations represent the final phase of infilling of the foreland basin and a return to a pattern of stable, outer carbonate platform sedimentation that persisted through the Llandovery (Anticostian). The Ellis Bay Formation, however, comprises an alternation of lithologies permitting recognition of seven members. Six of these were long recognized (Twenhofel 1928; Bolton 1972) and minor stratigraphical revision by Petryk (1979, 1981*a*) modified these to seven. This alternation has been interpreted as caused by eustatic sea-level changes associated with the Late Ordovician north African glaciation (McCracken & Barnes 1981*a*; Petryk 1981*b*; Johnson *et al.* 1981; Barnes 1986). The Ellis Bay Formation, redefined by Petryk (1979, 1981*a*) to extend only up to the level of the bioherms, is about 75 m


SYSTEM
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OSTRACODE ZONE
ARCITARCH ZONE
SILURIAN
LOWER
Wenlock
Llandovery
ORDOVICIAN
UPPER
Ashgill
Sheinwoodian
Telychian
C6
C5
C 3-4
C 1-2
Aeronian
B 1-3
A
Rhuddanian
2-4
A1
Hirnantian
Rawtheyan
Cautleyan
Pusgillian
Gwernfelen
Cerig
Wormwood
Rhydings
Trefawr
Crychan
Bronydd
Scrach
Tridwr
Jongian
Spirian
Leangian
Skinnerbukta
Vik
Rytteråker
Solvik
Langøyene
Husbergøya
Wenlock
Llandovery (Anticostian)
Cincinnatian
Jumpersian
Menierian
Gamachian
Richmondian
Maysvillian
Chicotte
Jupiter
mbr. 4
mbr. 1-3
Gun River
Becscie
Ellis Bay
m. 7
mbr. 1-6
Vauréal
Macasty
Pterospathodus amorphognath.
Icriodella inconstans
Ozarkodina aldridgei
D. staurognathoides
Icriodella discreta
—
Icriodella deflecta
O? nathani
13
Amorphognathus ordovicicus
12
11
A. superbus
Zygobolba decora
Zygobolba anticostiensis
Zygobolba erecta
Interval Zone
Jonesites semilunatus
Domasia symmetrica–D. trispinosa
Multiplicisph n. sp.–Eupoikilofusa stericula
Tunisphaeridium tentaculaferum–Multisphaer. n. sp.
Evittia birminghamensis–Leiofusa n. sp.
Baltisphaer. plicatispinae–B. verrucutum

Fig. 2 Chronostratigraphy, lithostratigraphy and biostratigraphy of Anticosti Island succession with correlations to the succession in Britain and Norway (from Barnes, in press).

thick. Members 1, 3 and 5 are more argillaceous than members 2, 4 and 6 and are more recessive; they consist dominantly of nodular, argillaceous limestone, mainly skeletal wackestone to packstone, with lenses of packstone to grainstone; interbeds and films of green and grey shale are common. These members are particularly fossiliferous with abundant brachiopods and common cephalopods, gastropods, trilobites, bivalves, aulacerid stromatoporoids, ostracodes, conodonts, and palynomorphs. Members 2 and 4 consist dominantly of thin- to medium-bedded limestone, mainly lime mudstone, with minor regular interbeds of grey shale; member 6 is a higher energy, cross laminated wackestone to packstone. Members 2, 4 and 6 are less fossiliferous than the other interbedded members, yielding sparse brachiopods, corals, aulacerids and microfossils. Member 7 consists of a basal oncolitic platform bed, 40 cm thick at Ellis Bay, which extends over most of the island and on which are developed small bioherms, typically 2 m high and 4–8 m wide (Figs 3, 4). These can be studied in vertical profile in the cliffs and in sequential horizontal profiles in the wave platform. Detailed stratigraphical descriptions of Ellis Bay Formation sections, particularly across the boundary interval, are given by Barnes *et al.* (1981) and McCracken & Barnes (1981*a*). The faunas of the Ellis Bay are abundant and diverse. In his pioneer study, Twenhofel (1928) described 172 species; later studies, particularly on microfossils not considered by Twenhofel, have probably doubled this figure. Twenhofel (1928) recognized that the Ellis Bay was of post-Richmondian age and proposed the term Gamachian for this latest Ordovician interval. This stage (Fig. 2) was largely ignored for half a century, but the recent Anticosti conodont work has demonstrated its validity as a North American regional stage (McCracken & Barnes 1981*a*; Barnes *et al.* 1981; Barnes, in press; McCracken & Nowlan, in press). Member 7 of the Ellis Bay Formation includes the Ordovician–Silurian boundary as defined on conodonts (McCracken & Barnes 1981*a*); the correlation of this level with the base of the *A. acuminatus* Zone at the Dob's Linn stratotype is discussed below.

The Becscie Formation was initially estimated at about 80 m thick by Twenhofel (1928) and Bolton (1972). Petryk (1979, 1981*a*) included most of Bolton's member 6 of the Ellis Bay Formation in the lower Becscie and his enlarged Becscie measures 131–173 m thick, with four informal members. The formation consists primarily of thin to thick bedded lime mudstone to bioturbated skeletal wackestone with brachiopod packstone and grainstone, intrarudstone, and some ball and pillow slump structures. In the upper third, packstone and grainstone are more prominent together with green shale. Much of the formation is extremely fossiliferous with concentrations of *Virgiana barrandei* (Billings) as well as corals, bryozoans and algae. Conodonts (McCracken & Barnes 1981*a*; Fåhraeus & Barnes 1981) and ostracodes (Copeland 1974) indicate an early Llandovery age (Rhuddanian; Menierian).

Above the Becscie lie the Gun River, Jupiter and Chicotte formations. These cover the middle to late Llandovery interval and are not part of this present paper. P. Copper has been studying the brachiopods of Anticosti (e.g. Copper 1977, 1981) and preliminary results of acritarch and chitinozoan studies have been published (Duffield & Legault 1981; Achab 1981).

There have been few detailed studies of the sedimentology of the Anticosti lithostratigraphical units. General reviews and interpretations have been given by Petryk (1981*a*) and in the several papers dealing with conodont faunas referred to above. Near the boundary, the sedimentology and palaeoecology of the bioherms, mainly from the eastern part of the island, was undertaken by Lake (1981). Orth *et al.* (1986) failed to detect any iridium anomaly across the Anticosti boundary interval that may have explained the systemic boundary extinctions through a bolide impact. Séguin & Petryk (1984) have produced some preliminary results of palaeomagnetic studies and J. Kirschvink and colleagues have recently begun a project to determine a possible magnetostratigraphic record in the sequence.

Palaeontology

Within the overall stratigraphy of the Anticosti sequence described above, consideration of the faunas and floras will be restricted here largely to the boundary interval.

Macropalaeontology

Graptolites. A separate paper by Riva (this volume, p. 221) reviews the Anticosti graptolite faunas.

Trilobites. Bolton (1981) reported and illustrated the most abundant and diverse of the Anticosti trilobite faunas which occurs in the upper member of the Vauréal Formation as the *Ceraurinus icarus* (Billings) Richmondian fauna. A less diverse fauna occurs in the Ellis Bay Formation and includes *Isotelus, Toxochasmops anticostiensis* (Twenhofel), *Otarion anticostiensis* (Twenhofel), with a member 7 interbiohermal association of *Primaspis* n. sp., *Cyphoproteus*(?) sp., *Calymene* sp. and *Amphilichas* sp. The boundary interval fauna is currently under study and preliminary results have been presented by Chatterton *et al.* (1983) and Lésperance (1985). They report that trilobite genera typical of the Ordovician disappear at the oncolitic platform bed, member 7 of the Ellis Bay Formation including *Celtencrinurus, Isotelus, Nahannia, Platycorphe* and *Toxochasmops.* The overlying 45 m of the lower Becscie Formation (of Petryk) does not contain diagnostic trilobites until the appearance of *Acernaspis.* Lespérance (1985) emphasizes the significance of this occurrence and infers a correlation with the *A. acuminatus* Zone. Barnes & Bergström (this volume), however, caution that its first appearance in Norway is higher, as could be its appearance on Anticosti.

Brachiopods. Lespérance (1985) has reviewed the boundary interval brachiopod data. *Vellamo*, a typical Ordovician genus, ranges up to 30 cm above the oncolitic platform bed, member 7, Ellis Bay Formation. As with the trilobites, the next 40 m of the lower Becscie contains few diagnostic brachiopods (e.g., *Parastrophinella reversa* in growth position; *Stricklandia* sp.). At about 100 m above the base of the Becscie is the first appearance of *Virgiana* sp., a level which Lespérance considers may be as low as the *A. acuminatus* Zone or *Cystograptus vesiculosus* Zone.

The distribution of the atrypoid brachiopods was reviewed by Copper (1981). Three species of *Spirigerina* occur in the Ellis Bay Formation and this genus is only known elsewhere in North America Ordovician strata from the Edgewood Group, Missouri (?Gamachian). Different forms of this genus, together with *Atrypina gamachiana* (Twenhofel), occur above the oncolitic platform bed which Copper (1981) considered as a suitable level for the systemic boundary. *Zygospiraella planoconvexa*, a typical Rhuddanian index fossil, occurs higher in the lower Becscie, below or at a level where *Virgiana* and the trilobite *Acernaspis* occur (e.g. Lespérance 1985: figs 3, 4).

Cocks & Copper (1981) reported a *Hirnantia* fauna from a thin interval, 4·5 m below the oncolitic platform bed, in eastern Anticosti. This level is about 5 m below the occurrence of Silurian conodonts at this locality (Nowlan 1982). Since no internal moulds were illustrated, Lespérance (1985) has queried the assignment of these brachiopods to the *Hirnantia* fauna, but recognized that this fauna does appear at an equivalent level to the south in the Gaspé region.

Other macrofossils. Although commonly abundant in the Anticosti sequence, insufficient work has been completed or published on other groups of macrofossils to add much resolution to defining the systemic boundary in this region. Aulacerid stromatoporoids range only into member 7, Ellis Bay Formation and are present in the oncolitic platform bed (Bolton 1981; Cocks & Copper 1981; Petryk 1982*c*). The global change from a labechiid to a clathrodictyid assemblage near the systemic boundary was documented by Webby (1980). The coral genus *Calapoecia*, typically regarded as Ordovician, occurs in the bioherms and up to 20 m above the base of the Becscie Formation of Petryk (Bolton 1981). Another such Ordovician genus, *Acidolites*, is also known to extend into the upper Becscie Formation (Bolton 1981) and the distribution of species on Anticosti, especially in the member 7 bioherms, has been documented by Dixon (1986). Some preliminary work on algae, including those in the bioherms, have been published by Copper (1977), Bolton (1981), and Gauthier-Coulloudon & Mamet (1981). Bolton (1981) reviewed the occurrence of echinoderms, molluscs, and bryozoans but none of these groups is sufficiently well documented to be of biostratigraphical value for the boundary interval.

Micropalaeontology

Microfossils have been systematically collected from all of the Anticosti succession and provide the most precise biostratigraphic control. Ostracodes were investigated initially, followed by extensive conodont work, and acritarch–chitinozoan studies are now in progress with much of this collecting being tied to the conodont samples.

Ostracodes. The Anticosti ostracode faunas have been documented by Copeland (1970, 1973, 1974, 1981, 1983) for the Anticosti sequence and a series of zones and subzones established (Fig. 2). Increasing faunal provincialism occurs with the Silurian faunas (Copeland & Berdan 1977). In broadest terms, two distinct faunas occur. An older, predominantly Ordovician, hollinacean fauna is developed through the Vauréal, Ellis Bay and the lower 35 m of the Becscie formations and is assigned to the *Jonesites semilunatus* Zone with ten subzones. Much of this fauna is replaced (e.g., extinction of the Tetradellidae and Eurychilinidae) abruptly by an endemic beyrichiacean zygobolbid fauna. However, this turnover is not precisely defined since there is a 10 m interval in the lower Becscie which yields only sparse undiagnostic ostracodes. The *Euprimitia gamachei* Subzone, the highest in the *Jonesites semilunatus* Zone, occurs in the lower 35 m of the Becscie Formation of Petryk. Copeland (1983) reported the distinctive Baltic species *Steusloffina cuneata*, considered to be of Ordovician age, from 6 m above the base of the Becscie Formation. The earliest Silurian zygobolbinid ostracodes occur about 40–50 m above the first occurrence of *Virgiana* and *Acernaspis* and 70 m above the first appearance of Silurian conodonts. Most of the ostracode distributions are plotted by member and/or formation by Copeland (1970, 1973, 1974) which limits the degree of resolution of ostracode biostratigraphy.

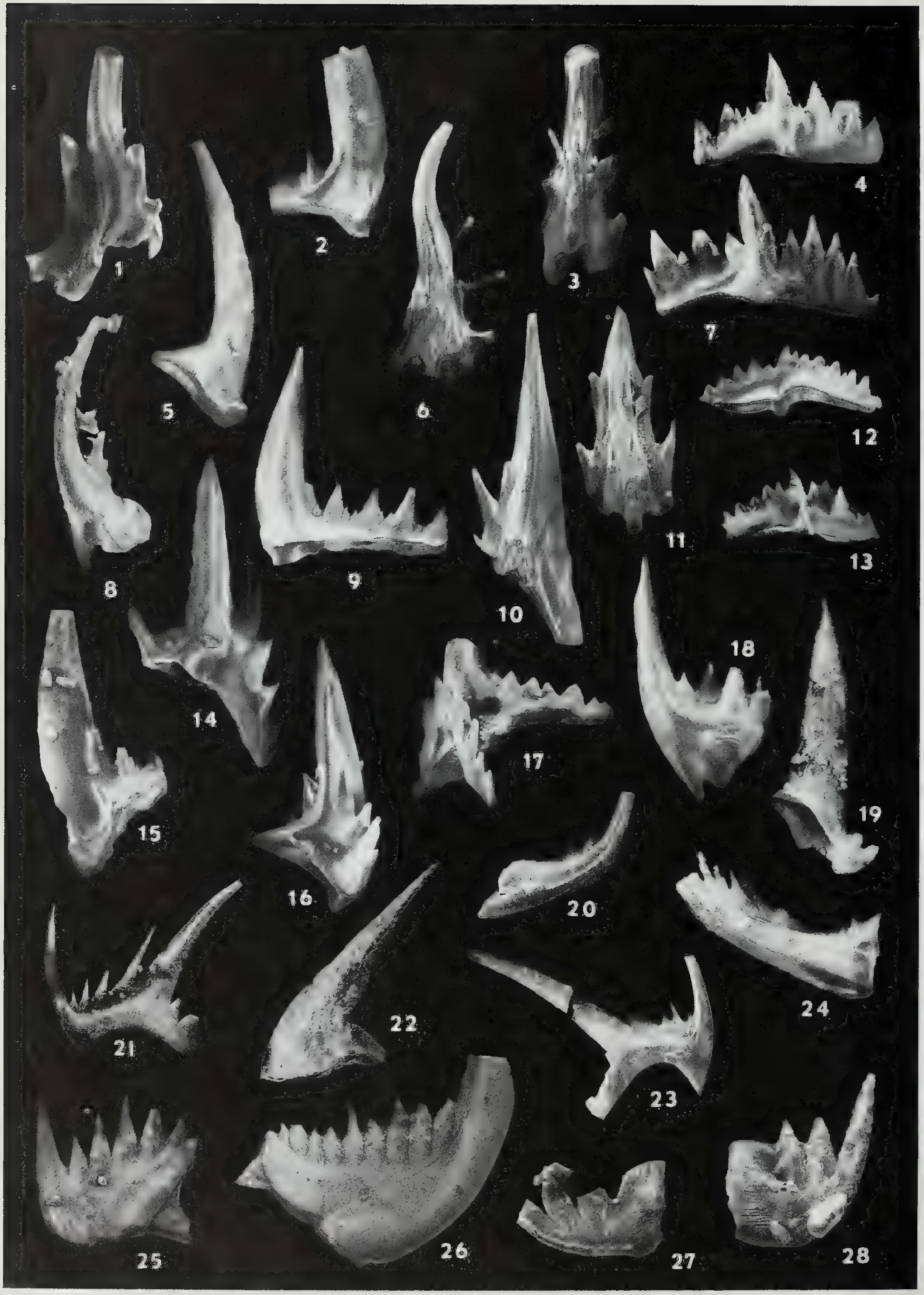
Palynomorphs. The chitinozoan faunas from the Vauréal and Ellis Bay formations have been described by Achab (1977*a*, *b*, 1981). A doctoral study of the latest Ordovician and the Silurian acritarchs was undertaken by Duffield (1982) and the preliminary results published (Duffield & Legault 1981). In both groups, significant turnovers occur at the level of the bioherms similar to that of the conodonts (see below).

For the chitinozoans, members 5 and 6 contain *Conochitina gamachiana* Achab, *C. micracantha* Eisenack and *C. taugourdeaui* Eisenack, which range up to the base of the bioherms. Above the bioherms, the fauna consists only of *Cyathochitina kuckersiana* Eisenack and *Ancyrochitina spongiosa* Achab with *Conochitina* sp. 1 of Achab higher in the Becscie.

The acritarch floral assemblage of the upper Ellis Bay Formation is of low diversity and abundance. Dominant taxa are *Baltisphaeridium plicatispinae* Gorka and *Multiplicisphaeridium* sp. 1 of Duffield & Legault. These taxa dominate up to the bioherms but the 2 m biohermal interval is generally barren of acritarchs. Some taxa range into the overlying Becscie but above the bioherms several new distinctive taxa appear including *Goniosphaeridium oligospinosum*, *Multiplicisphaeridium birminghamensis* and members of the *M. denticulatum* group. This diverse upper assemblage contains forms described elsewhere from Silurian strata in North America and Belgium.

Conodonts (Plates 1–3). The entire Anticosti outcrop was sampled at 2 m intervals for conodonts by Barnes and later expeditions have provided more intensive collections, particularly in the boundary interval. In all, some 700 samples have yielded over 150 000 conodonts. Most of the basic taxonomic and biostratigraphical results have now been published (McCracken & Barnes 1981*a*; Nowlan & Barnes 1981; Uyeno & Barnes 1983); for the upper Becscie–Gun River interval only preliminary results have appeared (Fåhraeus & Barnes 1981). These data have been important in a revision of North American chronostratigraphy using the Anticosti sequence as a reference section (Barnes & McCracken 1981; Barnes, in press) for the Gamachian, Menierian and Jumpersian stages (Fig. 2).

The Vauréal Formation yielded a diverse and particularly abundant conodont fauna of Richmondian age (Nowlan & Barnes 1981). The pattern of conodont communities reflects the gradually upward-shallowing sequences with *Phragmodus* and *Amorphognathus–Plectodina* dominated assemblages eventually being replaced by an *Oulodus–Aphelognathis* assemblage (Nowlan & Barnes 1981: figs 2, 3).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In the upper Vauréal a new distinctive genus, *Gamachignathus*, appears (McCracken *et al.* 1980) and then dominates the fauna of the entire Ellis Bay Formation, particularly the western sections. The Ellis Bay fauna contains many taxa ranging up from the Vauréal Formation but also new taxa such as *Aphelognathus* sp. aff. *A. grandis* and *Staufferella inaligera* as well as an absence of *Plectodina*. McCracken & Barnes (1981*a*) established conodont Fauna 13 for this Ellis Bay interval (following Faunas 1–12 of Sweet *et al.* (1971); see also Sweet (1984) for new conodont chronozones). This *Gamachignathus* fauna has since been recognized in other latest Ordovician marginal basins in North America, including the Matepedia Group, Gaspé (Nowlan 1981) and the Grog Brook Group, New Brunswick (Nowlan 1983), the Hanson Creek Formation, Ely Springs Dolomite, and Unnamed Limestone at Ikes Canyon, Toquima Range, Nevada and California (Ross *et al.* 1982: C11), the Fish Haven Dolomite of Utah (Leatham 1985), the Road River Formation of the Yukon (McCracken & Nowlan in press; McCracken & Lenz in press) and the Cape Phillips Group, Cornwallis Island, Canadian Arctic Archipelago (McCracken & Nowlan in press). This distinctive genus appears to have evolved in the latest Richmondian from *Birksfeldia* (Barnes & Bergström, this volume, p. 325).

McCracken & Barnes (1981*a*: fig. 12) have shown the distribution of nearly 40 form and multielement conodont species through the members of the Ellis Bay Formation. A remarkable turnover in the fauna occurs at the level of the bioherms. The Ordovician taxa range up to a level 50 cm above the oncolitic platform bed, that is in the lower 50 cm of the interbiohermal strata. At this level, taxa typical of the Silurian first appear (e.g. *Ozarkodina oldhamensis*). These intermingle with only a few taxa extending from underlying strata: *Gamachignathus ensifer*, *G. hastatus*, *Oulodus robustus* and the coniform taxa of *Panderodus*, *Pseudooneotodus*, *Decoriconus*, *Walliserodus* and *Staufferella*. Of these, *Gamachignathus* and *Staufferella* become extinct 1·5–2·0 m higher in the section, at the base of the Becscie Formation of Petryk. Within a few metres of the first appearance of Silurian conodonts, several other distinctive Silurian taxa appear including *Distomodus* sp. aff. *D. kentuckyensis*, *Icriodella discreta*, *I. deflecta*, *Oulodus*? *kentuckyensis*, *O.*? *nathani* and *Spathognathodus manitoulinensis*. The base of the Silurian on Anticosti was defined using conodonts as the first appearance of *Ozarkodina* (*O. hassi* and/or *O. oldhamensis*) (McCracken & Barnes 1981*a*; Barnes & McCracken 1981). These authors also

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PLATE 1 All figures × 70 except fig. 2 × 100, fig. 11 × 85 and figs 12, 13, and 17 × 35. Type specimens deposited in the Geological Survey of Canada, Ottawa; sample number given in parentheses after GSC type number.

Figs 1–8 *Gamachignathus hastatus* McCracken, Nowlan & Barnes. (1, 6) Posterior and inner lateral views of keislognathiform elements; GSC 84971, GSC 84976 (S-1). (2, 5) Inner lateral views of cyrtoniodiform elements; GSC 84972, GSC 84975 (S-1). (3) Posterior view of hibbardelliform element; GSC 84973 (S-1). (4, 7) Outer lateral and inner lateral views of modified prioniodiform elements; GSC 84974 (S-1), GSC 84977 (2B-2). (8) Outer lateral view of cordylodiform element; GSC 84978 (2B-3).

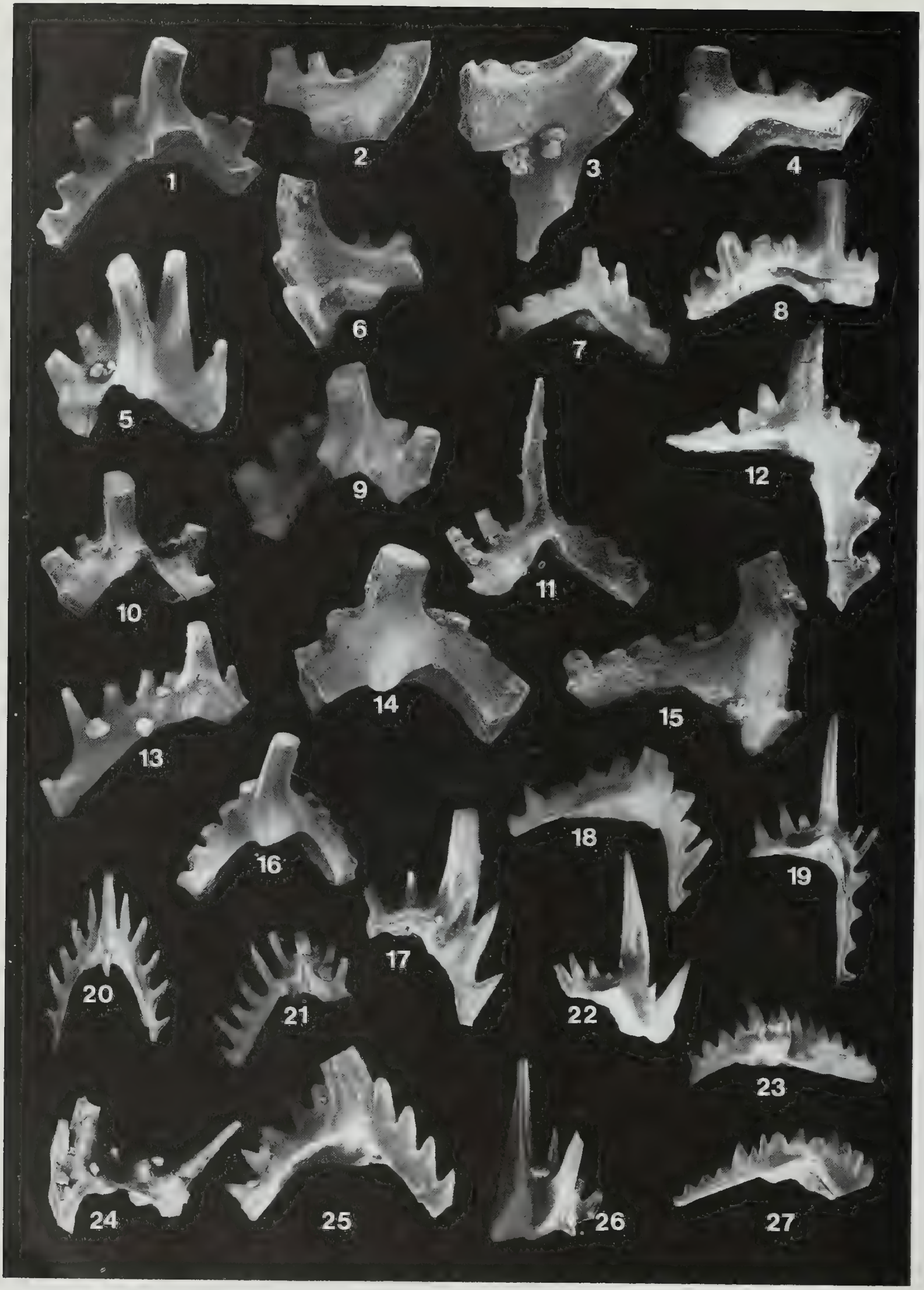
Figs 9–19 *Gamachignathus ensifer* McCracken, Nowlan & Barnes. (9) Inner lateral view of cyrtoniodiform element; GSC 84979. (10) Posterior view of keislognathiform element; GSC 84980. (11) Posterior view of hibbardelliform element; GSC 84981. (12, 13) Inner lateral and outer lateral views of modified prioniodiform elements; GSC 84982, GSC 84983. (14, 16, 17) Inner lateral, inner lateral and outer lateral views of prioniodiform elements; GSC 84984, GSC 84986, GSC 84987. (15, 18) Inner lateral and outer lateral views of cordylodiform elements; GSC 84985, GSC 84988. (19) Inner lateral view of falodiform element; GSC 84989. All specimens from sample S-1.

Figs 20, 24 *Pseudobelodina dispansa* (Glenister). (20) Lateral view of furrowed element; GSC 84990. (24) Lateral view of non-furrowed element; GSC 84994. Both specimens from sample S-1.

Figs 21–23 *Phragmodus undatus* Branson & Mehl. (21) Inner lateral view of trichonodelliform element; GSC 84991. (22) Outer lateral view of oistodiform element; GSC 84992. (23) Inner lateral view of cordylodiform–cladognathiform element; GSC 84993. All specimens from sample S-1.

Figs 25, 26 *Plegagnathus dartoni* (Stone & Furnish). (25) Outer lateral view of recurved element; GSC 84995 (S-145). (26) Inner lateral view of reclined element; GSC 84996 (S-1).

Figs 27, 28 *Pseudobelodina vulgaris vulgaris* Sweet. (27) Inner lateral view of broadly curved element; GSC 84997 (S-1). (28) Outer lateral view of tightly curved element; GSC 84998 (S-1).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established the *Oulodus*? *nathani* Zone for the earliest Silurian strata, lying below the *Distomodus kentuckyensis* Zone known elsewhere in North America (Fig. 2). In all the Anticosti conodont studies this conodont faunal turnover is by far the most profound and it is also a global event (Barnes & Bergström, this volume). In other carbonate sequences this same sharp boundary level can also be recognized. The *O.*? *nathani* Zone has been recognized elsewhere, for example in Gaspé, Quebec (Nowlan 1983) and the Oslo region of Norway (Aldridge & Mohamed 1982) based on the presence there of *O.*? cf. *O. nathani.*

The precise conodont faunal changes across the systemic boundary at the Ellis Bay and Salmon River sections, western and east-central Anticosti, were documented by McCracken & Barnes (1981*a*: figs 12, 14, tables 1–7). Cluster analysis was used to determine the changing community patterns with time, particularly with respect to east–west facies change. Additional collecting across the boundary interval was made by Duffield & Barnes in 1979 and the author in 1982 at Pointe Laframboise (Petryk 1981*a*: fig. 11), and west and east sides of Ellis Bay (Petryk 1981*a*: figs 12, 14) and at Salmon River (Petryk 1981*a*: figs 22, 23). These sections are described in both McCracken & Barnes (1981*a*) and Barnes *et al.* (1981).

The new conodont data are shown in Fig. 3 and Table 1. These three sections were closely sampled in each of these three sets of collections, resulting in sampling across the boundary interval at 10–20 cm intervals with each sampled interval being about 10 cm in thickness. In all, over 250 samples were taken through the 4–5 m interval at these three sections. The number of specimens per species per sample were tabulated by McCracken & Barnes (1981*a*) and Table 1 herein records similar data for the 1979 and 1982 collections. The latter two collections were taken close to the bioherms and produced much lower yields. Conodonts in general are rare in biohermal facies and to test this in the Anticosti sequence several samples (e.g. 2A.13–2A.15; 2B.14–2B.15) were taken from within the bioherms (Figs 3, 4B). All but one were barren and the exception contained only one specimen.

The faunal change occurring in this boundary interval described by McCracken & Barnes (1981*a*) and further by McCracken & Nowlan (in press), is substantiated in the new collections at each of the three sections. Some slight adjustments to the ranges of certain species can be noted. The general pattern is of an assemblage of Ordovician taxa up to the level of, and including, the oncolitic platform bed, member 7, Ellis Bay Formation, dominated by *Gamachignathus ensifer* and *G. hastatus.* At both the Pointe Laframboise and west side of Ellis Bay

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PLATE 2 All figures × 70 except figs 3, 6, 11, 17, 24, and 26 × 85, figs 7, 16, 18–23 and 27 × 35 and fig. 10 × 60. Sample numbers are as shown in Fig. 3, p. 211, except for S-143, 2 m below S-144; C-24, 1 m below oncolitic platform bed, east side Ellis Bay (Loc. 2C; Fig. 1).

Figs 1–3, 6–8 *Oulodus robustus* (Branson, Mehl & Branson). (1) Posterior view of zygognathiform element; GSC 85032 (2B-3). (2) Inner lateral view of cordylodiform element; GSC 85033 (2B-3). (3, 6). Inner lateral views of eoligonodiniform elements; GSC 85034, GSC 85037 (2B-3). (7) Outer lateral view of prioniodiniform element; GSC 85038 (C-24). (8) Posterior view of oulodiform element; GSC 85039 (2B-3).

Figs 4, 5, 9, 10, 12 *Oulodus ulrichi* (Stone & Furnish). (4) Inner lateral view of eoligonodiniform element; GSC 85035 (2B-3). (5, 9) Posterior views of zygognathiform elements; GSC 85036, GSC 85040 (2B-3). (10) Posterior view of trichonodelliform element; GSC 85041 (2B-3). (12). Posterior view of oulodiform element; GSC 85043 (S-143).

Figs 11, 13–19 *Oulodus rohneri* Ethington & Furnish. (11, 13) Posterior views of trichonodelliform elements; GSC 85042, GSC 85044 (2B-3). (14, 16) Posterior views of zygognathiform elements; GSC 85045, GSC 85047 (2B-3). (15) Inner lateral view of eoligonodiniform element; GSC 85046 (2B-3). (17) Inner lateral and posterior views of prioniodiniform element; GSC 85048 (S-143). (18, 19) Posterior view of oulodiform elements; GSC 85049, GSC 85050 (S-143).

Figs 20–27 *Aphelognathus* sp. aff. *A. grandis* Branson, Mehl & Branson. (20) Posterior view of trichonodelliform element; GSC 85051. (21, 26) Posterior views of zygognathiform elements; GSC 85052, GSC 85057. (22) Inner lateral view of cyrtoniodiform element; GSC 85053. (23, 27) Lateral views of aphelognathiform elements; GSC 85054, GSC 85058. (24) Inner lateral view of eoligonodiniform element; GSC 85055. (25) Inner lateral view of prioniodiniform element; GSC 85056. All specimens from sample S-143.

Table 1 Distribution of conodont species in the Ordovician–Silurian boundary interval, Anticosti Island, Québec. A: Pointe Laframboise (Locality 2A; Fig. 1). B: West side of Ellis Bay (Locality 2B; Fig. 1). C: 9 mile pool, Salmon River (Locality 5B; Fig. 1). Stratigraphical position of samples shown in Fig. 3 from collections by

Table 1A: Pointe Laframboise (Loc. 2A)

Species/Sample number	F3	F4	F5	F6	F7	F8	F9	F10	F11	F12	F13	F14	F15
Amorphognathus ordovicicus	—	—	—	—	—	—	—	—	—	—	—	—	—
Aphelognathus aff. *A. grandis*	—	—	—	—	—	—	—	—	—	—	—	—	—
Decoriconus costulatus	—	—	—	—	—	—	—	—	—	—	—	1	—
Drepanoistodus suberectus	—	—	—	—	—	—	—	—	—	—	—	—	—
Gamachignathus ensifer	—	23	3	9	1	7	6	3	—	—	—	—	—
G. hastatus	—	22	—	—	—	—	—	—	—	—	—	—	—
Oulodus robustus	—	25	—	—	—	—	—	—	—	—	—	—	—
O. rohneri	—	—	—	—	—	—	—	—	—	—	—	—	—
O. ulrichi	—	—	—	—	—	—	—	—	—	—	—	—	—
Panderodus spp.	—	—	1	1	—	—	—	—	—	2	186	76	8
Phragmodus undatus	—	—	—	—	—	—	—	—	—	—	—	—	—
Plegagnathus dartoni	—	—	—	—	—	—	—	—	—	—	—	—	—
Pseudobelodina dispansa	—	—	—	—	—	—	—	—	—	—	—	—	—
P. v. vulgaris	—	—	—	—	—	—	—	—	—	—	—	—	—
Pseudooneotodus beckmanni	—	3	1	—	—	—	—	—	—	—	—	2	—
Staufferella inaligera	—	—	—	—	—	—	—	—	—	—	—	—	—
Walliserodus cf. *W. curvatus*	—	2	2	1	—	—	—	—	—	—	—	—	—
Distomodus aff. *D. kentuckyensis*	—	—	—	—	—	—	—	—	1	2	—	—	—
Icriodella discreta	—	—	—	—	—	—	—	—	—	—	5	—	11
Oulodus? *kentuckyensis*	—	—	—	—	—	—	—	—	—	—	2	—	1
O.? *nathani*	—	—	—	—	—	—	—	—	—	—	—	—	—
Ozarkodina hassi	—	—	—	—	—	4	—	—	—	—	13	22	2
O. oldhamensis (+ramiforms of *O. hassi*)	—	—	—	—	6	3	5	5	—	—	72	56	2
Spathognathodus manitoulinensis	—	—	—	—	—	—	—	—	1	2	—	—	—
Walliserodus curvatus	—	—	—	—	—	—	—	—	—	—	38	31	—
Total specimens/sample	0	75	7	11	7	14	11	8	2	6	316	188	24

Duffield & Barnes, and Barnes; distribution data for other samples given in McCracken & Barnes (1981*a*). Average sample weight is 2 kg.

2A-1	2A-2	2A-3	2A-4	2A-5	2A-6	2A-7	2A-8	2A-9	2A-10	2A-11	2A-12	2A-13	2A-14	2A-15
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
1	5	2	—	—	—	—	—	1	4	—	—	—	—	—
—	—	—	2	—	—	—	—	—	—	—	—	—	—	—
—	5	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	1	2	—	1	—	—	—	6	—	1	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
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—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
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—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
1	10	3	4	2	1	1	0	2	13	1	9	0	0	1

Table 1B: West side, Ellis Bay (Loc. 2B)

Species/Sample number	H17	H18	H19	H20	H21	H22	H23	2B-1	2B-2
Amorphognathus ordovicicus	—	—	—	—	—	—	—	—	—
Aphelognathus aff. *A. grandis*	—	—	—	—	—	—	—	—	—
Decoriconus costulatus	—	—	—	—	7	—	—	—	—
Drepanoistodus suberectus	—	—	—	—	—	—	—	—	—
Gamachignathus ensifer	34	6	6	—	—	—	—	—	2
G. hastatus	—	—	—	—	—	—	—	—	4
Oulodus robustus	23	—	—	—	—	—	—	—	2
O. rohneri	—	—	—	—	—	—	—	—	—
O. ulrichi	—	—	—	—	—	—	—	—	—
Panderodus spp.	—	—	—	3	161	5	22	—	1
Phragmodus undatus	—	—	—	—	—	—	—	—	—
Plegagnathus dartoni	—	—	—	—	—	—	—	—	—
Pseudobelodina dispansa	—	—	—	—	—	—	—	—	—
P. v. vulgaris	—	—	—	—	—	—	—	—	—
Pseudooneutodus beckmanni	8	—	—	—	54	—	1	—	—
Staufferella inaligera	—	—	—	—	—	—	—	—	—
Walliserodus cf. *W. curvatus*	—	—	—	—	—	—	—	—	—
Distomodus aff. *D. kentuckyensis*	—	—	—	—	—	—	1	—	—
Icriodella discreta	—	—	—	2	9	1	16	—	—
Oulodus? *kentuckyensis*	—	—	—	—	1	—	—	—	—
O.? *nathani*	—	—	—	—	—	—	—	—	—
Ozarkodina hassi	—	4	4	—	—	—	14	—	—
O. oldhamensis (+ramiforms of *O. hassi*)	—	—	—	—	26	—	—	—	—
Spathognathodus manitoulinensis	—	—	—	—	—	—	—	—	—
Walliserodus curvatus	—	—	—	—	24	—	—	—	—
Total specimens/sample	65	10	10	5	282	6	54	0	9

(2B-3 & 2B-4)	2B-5	2B-6	2B-7	2B-8	2B-9	2B-10	2B-11	2B-12	2B-13	2B-14	2B-15
2	—	—	1	—	—	—	—	—	—	—	—
—	—	—		—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—
1	1	1	—	—	—	—	1	1		—	—
32	—	—	—	—	1	—	—	—	—	—	—
8	—	—	—	—	—	—	—	—	—	—	—
5	—	—	—	—	—	—	—	—	—	—	—
4	—	—	—	—	—	—	—	—	—	—	—
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1	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	1	—		—
—	—	—	1	1	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	1	3	—	—
—	—	—	—	—	—	—	—	—	1	—	—
—	—	—	—	—	—	1	1	—	—	—	—
—	—	—	—	—	—	1	—	—	—	—	—
—	—	—	—	—	—	1	—	—	2	—	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	3	—	—
53	1	1	2	1	1	3	2	5	39	0	0

Table 1C: Salmon River (Loc. 5B)

Species/Sample number	S144	S145	S146	S147	S148	S149	S150	S151	S152	S153	S1	S2	S3	S4	S5	S6	S7	S8
Amorphognathus ordovicicus	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Aphelognathus aff. *A. grandis*	—	17	—	—	—	—	—	—	—	—	38	—	—	—	—	—	—	—
Decoriconus costulatus	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Drepanoistodus suberectus	—	6	—	—	—	—	—	—	—	—	15	—	—	—	—	—	—	—
Gamachignathus ensifer	10	25	10	—	—	—	—	—	—	—	39	3	13	23	—	—	13	—
G. hastatus	1	5	—	—	—	—	—	—	—	—	34	—	—	—	—	—	—	—
Oulodus robustus	—	—	—	—	—	—	—	—	—	—	—	—	—	4	—	—	—	—
O. rohneri	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
O. ulrichi	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Panderodus spp.	6	19	3	10	—	5	200+	18	200+	54	98	11	6	—	—	—	—	—
Phragmodus undatus	2	1	—	—	—	—	—	—	—	—	3	—	—	—		—	—	—
Plegagnathus dartoni	—	1		—	—	—	—	—	—	—	1	—	—	—	—	—	—	—
Pseudobelodina dispansa	—	1	—	—	—	—	—	—	—	—	6	—	—	—	—	—	—	—
P. v. vulgaris	—	1	—	—	—	—	—	—	—	—	5	—	—	—	—	—	—	—
Pseudooneotodus beckmanni	—	3	2	—	—	1	45	6	30	11	2	—	—	—	—	—	—	—
Staufferella inaligera	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Walliserodus cf. *W. curvatus*	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Distomodus aff. *D. kentuckyensis*	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Icriodella discreta	—	—	—	—	—	3	—	—	1	—	—	—	—	—	—	—	—	—
Oulodus? kentuckyensis	—	—	—	—	—	3	—	—	—	1	—	—	—	—	—	—	1	—
O.? nathani	—	—	—	—	—	—	2	—	—	1	—	—	—	—	—	—	1	2
Ozarkodina hassi	—	—	—	—	—	2	4	—	—	1	—	—	—	6	—	10	—	—
O. oldhamensis (+ramiforms of *O. hassi*)	—	—	3	1	—	—	30	3	57	7	—	—	—	7	—	10	12	—
Spathognathodus manitoulinensis	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	1	4
Walliserodus curvatus	—	—	—	10	—	4	16	—	15	5	—	—	—	—	—	—	—	—
Total specimens/sample	19	79	18	21	0	18	297+	27	303+	80	241	14	19	40	0	20	28	6

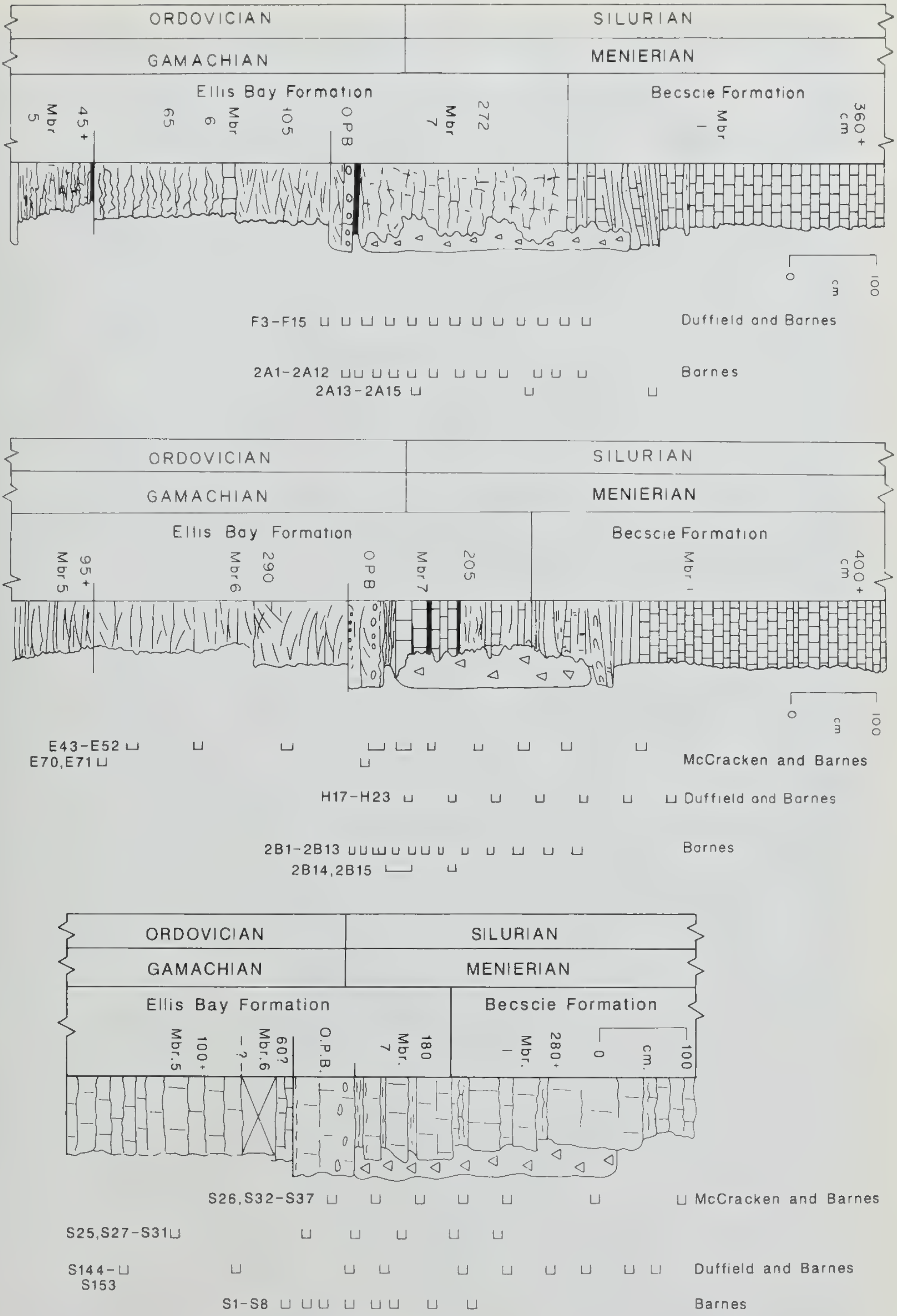
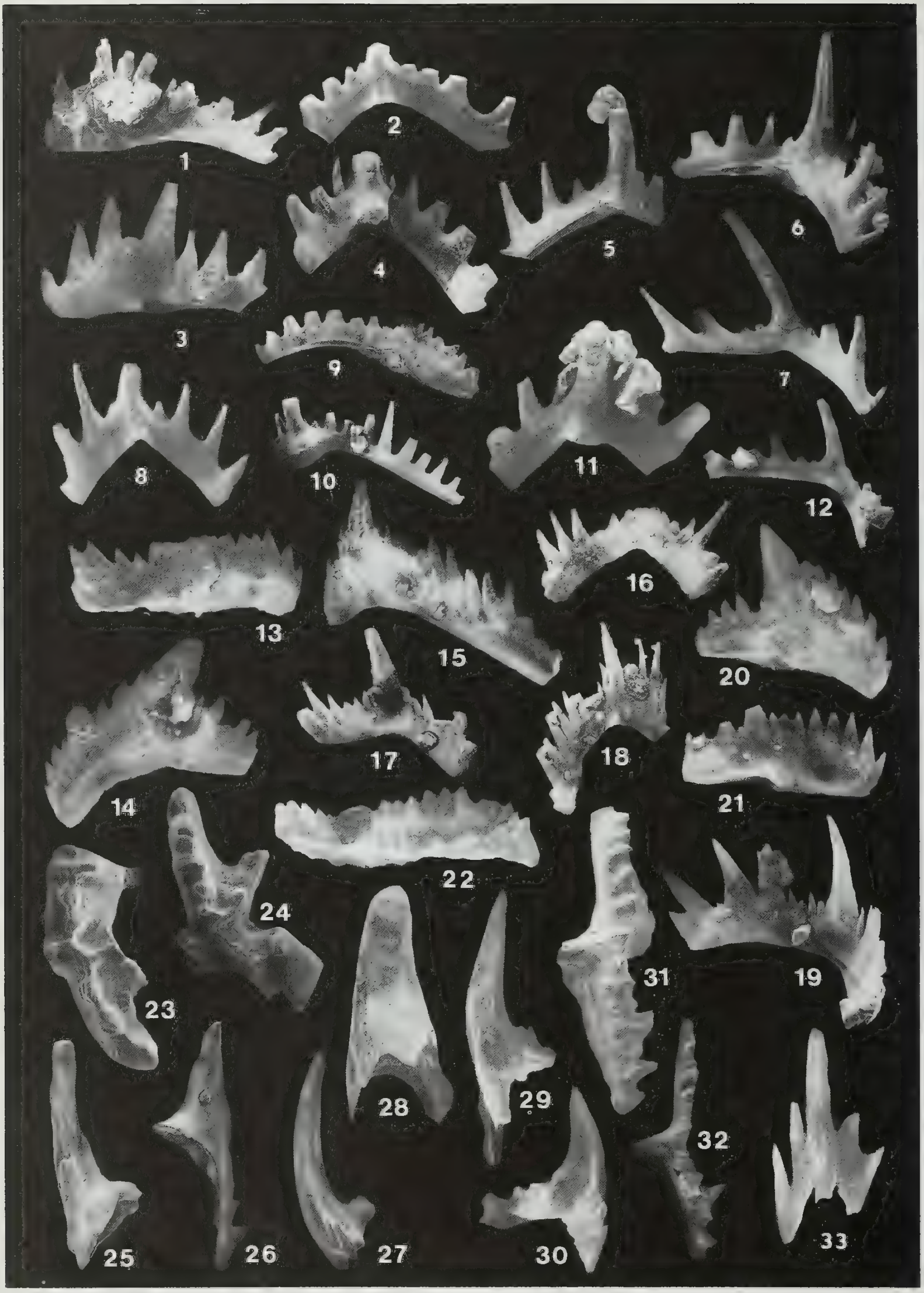

ORDOVICIAN
SILURIAN
GAMACHIAN
MENIERIAN
Ellis Bay Formation
Becscie Formation
Mbr 5
45+
65
Mbr 6
105
OPB
Mbr 7
272
Mbr 1
360+ cm
0
cm
100
F3–F15
Duffield and Barnes
2A1–2A12
2A13–2A15
Barnes
ORDOVICIAN
SILURIAN
GAMACHIAN
MENIERIAN
Ellis Bay Formation
Becscie Formation
Mbr 5
95+
Mbr 6
290
OPB
Mbr 7
205
Mbr 1
400+ cm
0
cm
100
E43–E52
E70,E71
McCracken and Barnes
H17–H23
Duffield and Barnes
2B1–2B13
2B14,2B15
Barnes
ORDOVICIAN
SILURIAN
GAMACHIAN
MENIERIAN
Ellis Bay Formation
Becscie Formation
Mbr.5
100+
–?–
Mbr.6
60?
O.P.B.
Mbr. 7
180
Mbr. 1
280+
0
cm.
100
S26,S32–S37
McCracken and Barnes
S25,S27–S31
S144–S153
Duffield and Barnes
S1–S8
Barnes

Fig. 3 Detail of lithostratigraphy of the Ordovician–Silurian boundary interval at Pointe Laframboise (left), west side of Ellis Bay (centre) and Salmon River (right) (Localities 2A, 2B, and 8B respectively on Fig. 1) showing position of conodont samples collected by McCracken & Barnes, Duffield & Barnes, and Barnes. Conodont distribution data from these samples given by McCracken & Barnes (1981*a*) and in Table 1 herein. OPB = oncolitic platform bed.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sections (Localities 2A, 2B, on Fig. 1), the first Silurian taxa (*Ozarkodina oldhamensis*, *O. hassi*, *Spathognathodus manitoulinensis* and *Oulodus*? *nathani*) appear about 90 cm above the base of the oncolitic platform bed, that is 50 cm above the top of this bed within the interbiohermal strata (Fig. 4D). At the Salmon River section (Locality 5B on Fig. 1; Fig. 4F), the later collections show the first occurrence to be still 90 cm above the base of the oncolitic platform bed but since this bed has thickened to 90 cm, from 40 cm in the western sections, the top 10 cm of this bed have now yielded Silurian taxa (Table 1). This is about 50 cm lower than the level reported by McCracken & Barnes (1981*a*) and perhaps the level reported by Nowlan (1982) from a coastal section further to the east. In the three sections, *Gamachignathus*, *Oulodus robustus* and *Staufferella inaligera* range through the next two metres, mixed with the early Silurian forms. At a level approximating to the base of Petryk's Becscie Formation (typically 2 m above the base of the oncolitic platform bed, and equivalent to a level within a metre of the top of the bioherms) these residual Ordovician taxa disappear and the earlier Silurian taxa are joined by other Silurian forms such as *Icriodella discreta*, *Icriodella deflecta*, *Distomodus* sp. aff. *D. kentuckyensis* and *Oulodus*? *kentuckyensis*.

Biostratigraphical correlations

This paper has reviewed the sequence of faunas through the systemic boundary interval on Anticosti and added new conodont data. Many of the references noted above include sections on the regional biostratigraphical correlations. Space limitations prevent a comprehensive dis-

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PLATE 3 All figures × 70 except figs 4–8, 18, 20, 21 × 85 and fig. 31 × 35. Sample numbers are as shown in Fig. 3, p. 211, except for S-154, S-155, 2 and 1·5 m above S-153; C-38 in Lower Becscie, 1·2 km east of Cap à l'Aigle (Loc. 3B; Fig. 1); F-16 is 2 m above F-15.

Figs 1–8 *Oulodus*? *nathani* McCracken & Barnes. (1, 3) Inner lateral views of modified oulodiform elements; GSC 84999, GSC 85001. (2) Posterior view of trichonodelliform element; GSC 85000. (4, 8) Posterior view of zygognathiform elements; GSC 85002, GSC 85006. (5, 6) Inner lateral views of lonchodiniform elements; GSC 85003, GSC 85004. (7) Inner lateral view of ligonodiniform element; GSC 85005. All specimens from sample S-154 except (1) which is from sample C-38.

Figs 9–12 *Oulodus*? *kentuckyensis* (Branson & Branson). (9) Lateral view of modified oulodiform element; GSC 85007 (F-15). (10) Lateral view of eupriodiodiniform element; GSC 85008 (S-153). (11) Posterior view of zygognathiform element; GSC 85009 (S-154). (12) Inner lateral view of ligonodiniform element; GSC 85010 (S-154).

Figs 13, 14 *Ozarkodina oldhamensis* (Rexroad). (13) Lateral view of spathognathodiform element; GSC 85011 (S-155). (14) Inner lateral view of ozarkodiniform element; GSC 85012 (S-155).

Figs 15–19 Ramiform elements of *O. oldhamensis* and *O. hassi* complex. (15) Lateral view of synprioniodiniform element; GSC 85013. (16) Posterior view of zygognathiform element; GSC 85014. (17, 19) Inner lateral views of ligonodiniform elements; GSC 85015, GSC 85017. (18) Posterior view of trichonodelliform element; GSC 85016. All specimens from sample S-155.

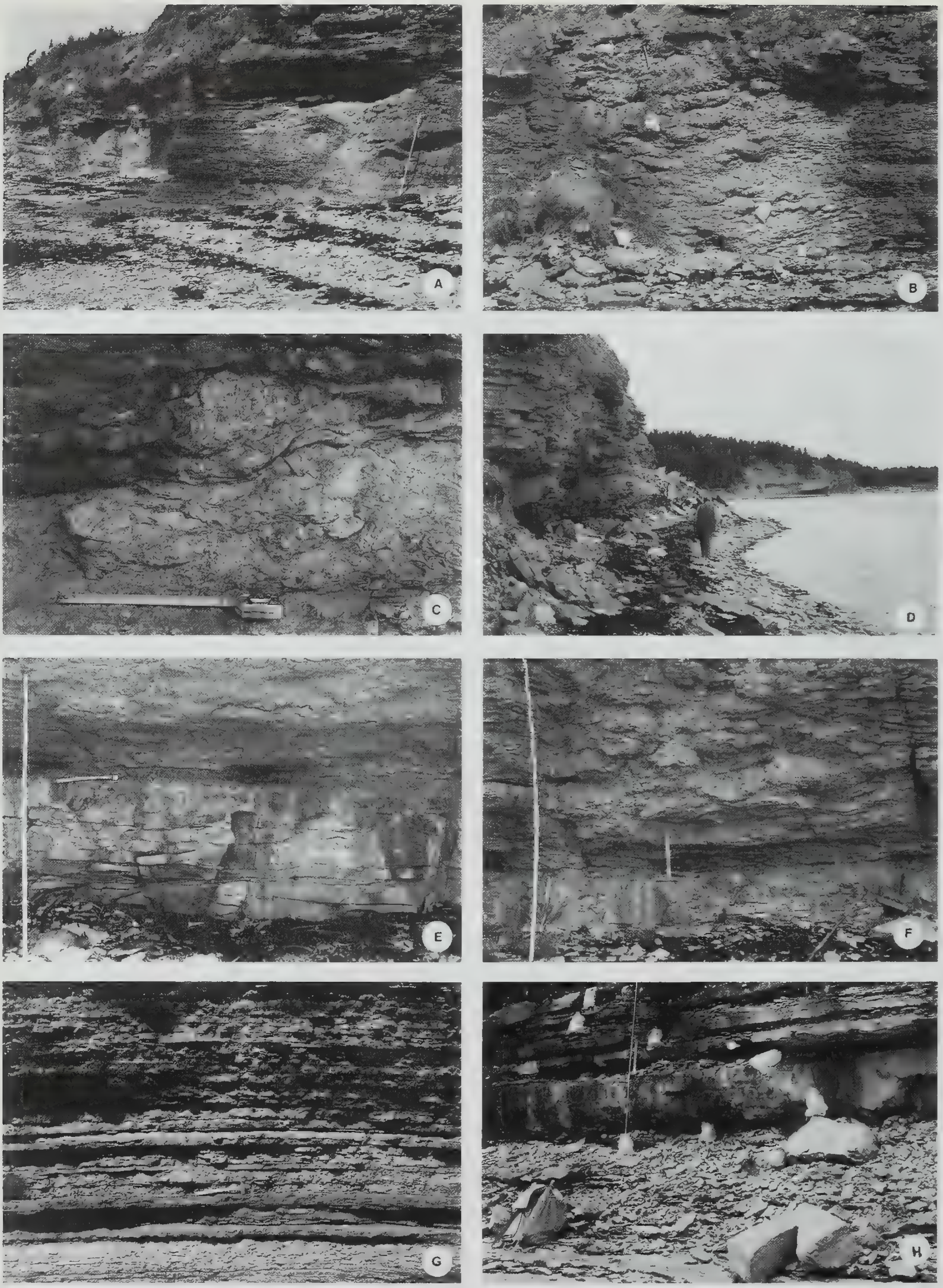
Figs 20, 21 *Ozarkodina hassi* (Pollock, Rexroad & Nicholl). (20) Inner lateral view of ozarkodiniform element; GSC 85018 (S-153). (21) Lateral view of spathognathodiform element; GSC 85019 (2A-10).

Fig. 22 *Spathognathodus manitoulinensis* Pollock, Rexroad & Nicholl. Inner lateral view of spathognathodiform element; GSC 85020 (S-8).

Figs 23–28 *Distomodus* sp. aff. *D. kentuckyensis* Branson & Branson. (23, 24) Upper view of platform elements; GSC 85021, GSC 85022. (25) Inner lateral view of distomodiform element; GSC 85023. (26) Inner lateral view of modified ambalodiform element; GSC 85024. (27) Inner lateral view of eoligonodiniform element; GSC 85025. (28) Posterior view of zygognathiform element; GSC 85026. All specimens from sample F-16.

Figs 29–31, 33 *Icriodella discreta* Pollock, Rexroad & Nicholl. (29) Outer lateral view of sagittodontiform element; GSC 85027 (2B-13). (30) Inner lateral view of ambalodiform element; GSC 85028 (2B-13). (31) Upper view of icriodelliform element; GSC 85029 (2B-12). (33) Posterior view of trichonodelliform element; GSC 85031 (2B-13).

Fig. 32 *Icriodella deflecta* Aldridge. Upper view of icriodelliform element; GSC 85030 (C-55: base of Gun River Formation, Locality 5C on Fig. 1).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cussion here of the correlations suggested by all the different fossil groups. Fairly precise correlations can be made from Anticosti to the various sections in Gaspé and New Brunswick, to sections in central and western North America, and to Norway (e.g. Lespérance 1985; Barnes & Bergström, this volume, p. 325). These correlations can be effected best through use of conodonts, ostracodes, shelly fossils and palynomorphs (Fig. 2).

The more difficult correlation is with oceanic graptolitic sequences, for example with the Dob's Linn stratotype. This problem has been addressed from different viewpoints by Barnes & Bergström; Barnes & Williams; and Riva (all in this volume). There is no precise correlation since Dob's Linn contains few fossils other than graptolites and these are rare in the Anticosti boundary interval. Barnes & Bergström (this volume) conclude that the conodont faunal turnover, so dramatically seen on Anticosti and elsewhere, must occur at a level within the upper *Glyptograptus persculptus* Zone up to, but no higher than, the base of the *Akidograptus acuminatus* Zone (the formally defined base of the Silurian). The major extinction event in conodonts and graptolites thus occurs within latest Ordovician time and not at the new systemic boundary. The earliest Silurian conodonts on Anticosti referred to in this paper may therefore be of latest Ordovician age (e.g. latest *G. persculptus* Zone) but at this point it is both impossible to be so precise and impractical not to refer them to the Silurian, since they are so distinctively different from Ordovician forms and form the basis for correlation in the extensive Lower Silurian carbonate platform sequences.

The conodonts, palynomorphs, aulacerid stromatoporoids, and, to a lesser extent, the brachiopods and trilobites show distinct faunal changes at essentially the same level in member 7 of the Ellis Bay Formation. Some other groups, however, seem to show a significant change within 20–50 m higher in the sequence (e.g. ostracodes, corals). Assuming that the extinctions are induced directly or indirectly by the glacial climatic events (e.g. Brenchley 1984; Barnes 1986) it is to be expected that different fossil groups would respond to such environmental pressures in different ways and at slightly different times.

Future studies and potential

The beauty of Anticosti Island is not only in its modern fauna, flora and scenery but in the magnificent quality and potential of the stratigraphical sections. The Ellis Bay section has virtually all the attributes for a boundary stratotype: well exposed, continuous sedimentation, variable lithology, abundant and diverse faunas and floras, no structural deformation, little thermal alteration, geographically accessible, a reasonably sound knowledge base and long-term protection. In comparison to Dob's Linn, it lacks abundant graptolites and historical precedence. However, in most of the other criteria, the Dob's Linn section has serious weaknesses to the point at which important stratigraphical principles have been disregarded or overruled in making the final stratotype decision (Lespérance *et al.* 1987). Whereas there may be little more significant faunal data to be extracted from the well-collected Dob's Linn section,

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Fig. 4 Ordovician–Silurian boundary interval, Anticosti Island, Québec. A–C, Point Laframboise (Locality 2A, fig. 1); A: 2 m tape is at level of bioherms, member 7, Ellis Bay Formation overlain by lower Becscie Formations; B: detail of biohermal–interbiohermal relationships, grainstones well developed against upper quarter of bioherm; C: detail of upper bioherm surface with crinoidal grainstones abutting and overlapping coral heads. D–G, West side of Ellis Bay, north of Cap Henri (Locality 2B, Fig. 1); D: view of cliff exposures of member 5, Ellis Bay Formation (background) and member 6 (foreground), wave platform covered by high tide; E: members 6 and 7, Ellis Bay Formation, hammer (40 cm) rests on top of oncolitic platform bed which forms base of member 7, overlain by this recessive shale and interbiohermal strata; F: similar view to E but showing bioherm developed on oncolitic platform bed above hammer; systemic boundary drawn 50 cm above top of oncolitic platform bed; G: lower Becscie Formation, hammer is 40 cm. H, Salmon River, 9 mile pool (Section 8B, Fig. 1); massive bed in centre is 90 cm oncolitic platform bed, member 7, Ellis Bay Formation, overlain by interbiohermal strata; hammer is 40 cm, earliest Silurian conodonts occur in top 10 cm of massive bed.

the Anticosti sequence will clearly continue to yield a wealth of new data and its future potential in studies through the Ordovician–Silurian boundary is probably unsurpassable. Although the systemic boundary has been decided, its reconsideration may be necessary (Lespérance *et al.* 1987). Future work will also concentrate on unravelling the type and timing of processes that caused such major extinctions. Sepkoski (1982) has calculated that 22 per cent of all families became extinct at this boundary, making it second only to the terminal Permian extinction in severity among Phanerozoic biotic crises.

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Graptolites at and below the Ordovician–Silurian boundary on Anticosti Island, Canada

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Synopsis

Graptolites in the lower and middle Vauréal Formation of Anticosti Island, Canada, form a discrete assemblage renamed the *Amplexograptus prominens* Zone, characterized by *Amplexograptus latus*, *Rectograptus abbreviatus*, *Amplexograptus prominens* and *Paraclimacograptus decipiens* sp. nov.: these suggest correlation with the *Dicellograptus anceps* Zone of Scotland, the *Climacograptus pacificus* Zone of north-eastern Siberia and Kazakhstan, and the Wufeng Shale of Central china. Graptolites are rare in the upper Vauréal Formation. A few have been collected from the upper members of the Ellis Bay Formation and the lower members of the Becscie Formation, but not in sufficient numbers to form a zonal assemblage. Most of them belong to the *normalis* group for which the new genus *Scalarigraptus* is proposed. The most common graptolite is *Scalarigraptus angustus*, which is known to range through the upper Ashgill and the lower Llandovery Series. Two fragmentary specimens identified as *Rectograptus abbreviatus* have been collected from the top (Member 6) of the Ellis Bay Formation. This species is only known from the Upper Ordovician and may be taken to indicate that the top members (6 and 7) of the formation belong to the Ordovician System.

Introduction

In an earlier paper Riva (Riva & Petryk 1981) reviewed and updated the work done by previous workers on graptolites from the Island of Anticosti, either as part of a general palaeontological study (Twenhofel 1928) or as detailed morphological studies of isolated graptolites (Barrass 1953; Strachan 1954). It also updated the study of subsurface collections which had been extracted by Riva (1969) from three drill cores during the summers of 1964 and 1965, and presented an evaluation of 33 new collections made by A. A. Petryk from 1975 to 1979 from the upper Vauréal to the Jupiter Formations. An accompanying range chart showed the stratigraphical position of all graptolites hitherto identified from surface collections. This chart will undergo further revisions and refinements as new morphological studies and revisions of type collections are made known. Part of this work is incorporated into this paper together with data on new collections made by Petryk from 1981 to 1983.

This paper is primarily concerned with the graptolites collected at or just above or below the Ordovician–Silurian boundary now located at the Ellis Bay–Becscie formational contact (Fig. 1) (Lespérance 1985). It also re-evaluates the fauna of the *Amplexograptus prominens* Zone of the lower and mid-Vauréal Formation and correlates it with the zonal successions of Scotland, the U.S.S.R., China and Australia. Figure 1 shows the range of all graptolites hitherto identified from the mid-Vauréal to the lower Becscie Formations plotted against the revised surface stratigraphy and nomenclature of Petryk (1979). The graptolites from below the mid-Vauréal Formation, which are known only from drill-cores, have been treated separately (Riva 1969).

A graptolite zone and other graptolites

The *Amplexograptus prominens* Zone. This is the youngest of the zones proposed by Riva (1969) from his study of drill cores and the only one recognized from surface exposures of the Vauréal Formation. In the N.A.C.P. well (Riva 1969: fig. 12) it spans much of the lower Vauréal between the 2047–1734 ft level (614–412 m), for a thickness of 202 m. In both the N.A.C.P. and the L.G.P.L. wells (Riva 1969: figs 11 and 12) it follows on the *Dicellograptus complanatus* Zone which spans most of the underlying 'English Head' Formation (to be renamed the Princeton Lake Formation) for a thickness of 193 m. Originally, Riva (1969) named the *A. prominens* Zone

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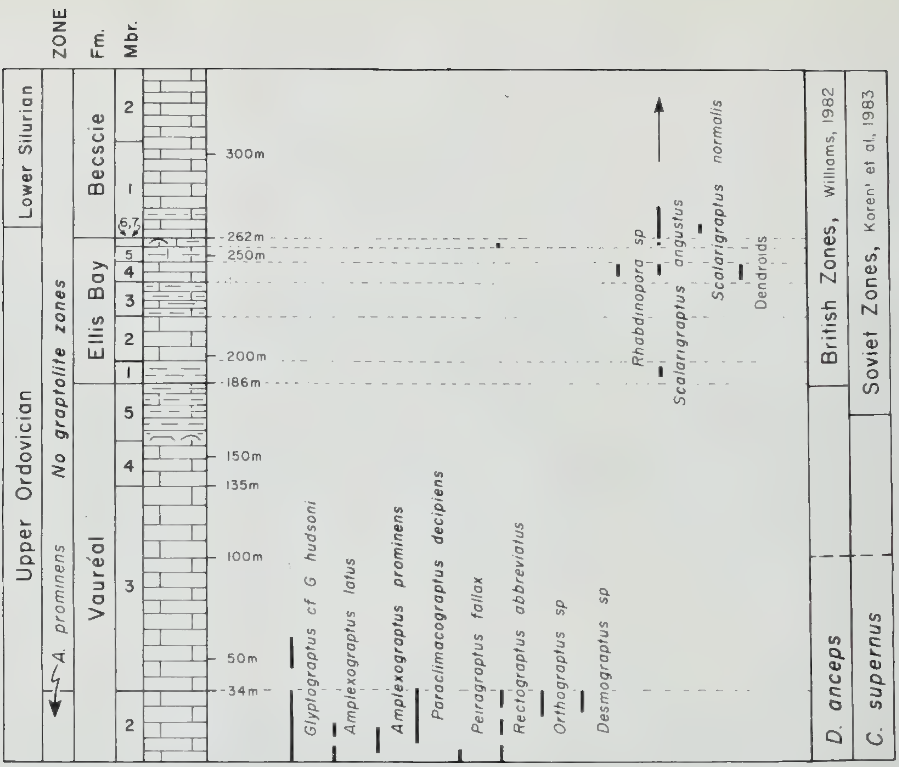


Fig. 1 Graptolite ranges in the upper Vauréal, Ellis Bay and lower Becscie Formations of Anticosti Island.

the *Climacograptus prominens–elongatus* Zone and interpreted its fauna (constituted primarily of biserial graptolites not easily related to those of other successions) as representing a level 'intermediate between ... the youngest Ordovician and the oldest Silurian' (1969: 551). He also referred the species used to name the zone to *Climacograptus* rather than *Amplexograptus*, as Barrass (1953) had done, because most specimens recovered from the core possessed climacograptid thecae with everted apertures rather than amplexograptid thecae. In 1981 he renamed the zone the *Amplexograptus inuiti* Zone on the recognition that *A. elongatus* Barrass was identical to *Amplexograptus inuiti* described by Cox (1933) and also its junior synonym. He also re-interpreted *Amplexograptus prominens* Barrass as a subspecies of *A. inuiti.*

In 1985, I studied and sorted out the type material of *Climacograptus latus* Elles & Wood and came to the conclusion that this species belongs to *Amplexograptus* rather than *Climacograptus, s.l.*, and is also identical to, and the senior synonym of, *A. inuiti. A. latus* was erected on flattened, fragmentary material and *A. inuiti* (Figs 4b–c) on excellent, isolated specimens from Akpatok Island in northeastern Canada. Cox (1933: 2) pointed out the similarity of *A. inuiti* to *A. latus*, but refrained from considering the two species identical because the thecal apertures of *A. latus* were 'more even' and lacked genicular flanges. In reviewing the type specimens of *A. latus*, I recognized apertural lappets in all specimens retained in the species and also residual genicular flanges (Figs 2a–h), but not in the specimens that I have excluded from it (Figs 2i–j), which belong to *Climacograptus tubuliferus* Lapworth. These features are even more pronounced in the topotype material recently identified as *C. latus* by Williams (1982: pl. 3, figs 12–18). The occurrence of *A. latus* in the *A. prominens* Zone of Anticosti is critical, for it allows us to correlate this zone with the *Dicellograptus anceps* Zone of Scotland and the *Cli-*

macograptus pacificus Zone of the U.S.S.R. and their equivalents in China, the Yukon, and elsewhere, something which had not hitherto been possible. I have, however, refrained from naming this zone after either *D. anceps* or *C. pacificus* because neither graptolite has been recovered from Anticosti.

Amplexograptus prominens itself is morphologically quite distinct from *Amplexograptus latus* and cannot be regarded as a mere subspecies or a morphological variant of it. The study of an original collection of *A. prominens* (made by Col. C. C. Grant) from the type strata at Observation Cliff on the north shore of Anticosti Island fully confirms Barrass' (1953) original diagnosis of this species. *A. prominens* is characterized by broad, short rhabdosomes which expand rapidly from a narrow proximal end (first pair of thecae), by prominent genicular flanges and the absence of a mesial spine on th 1^1. The long genicular flanges and the lack of a mesial spine on th 1^1 set *A. prominens* well apart from all other species of *Amplexograptus*, although it shares with them a similar type of proximal-end development (early prosoblastic) and thecal style (amplexograptid with well-developed lappets) (Riva 1987). *A. prominens* is a unique species, known up to now only from the upper Vauréal Formation of Anticosti. It is the last *Amplexograptus*. It could well be the immediate ancestor to *Paraclimacograptus decipiens* sp. nov. which has a long range through the upper Vauréal and with which it has been confused in the past. *P. decipens* differs from *A. prominens* both in thecal form and the nature of genicular ornaments (Fig. 2s) but otherwise it shares with it the same type of proximal development and general distal rhabdosome structure (Figs 2o–r). On the other hand, the isolated specimens from Manitoba referred to *A. prominens* by Jackson (1973: 2–4; text-figs 2B, E and F) are close to the topotypes of the older *Paraclimacogratus manitoulinensis* (Caley) shown here as Figs 5g, h and i. Occasional low or incipient lappets are present both on the everted thecal apertures of the Manitoba specimens and the topotype specimens of *P. manitoulinensis*, and the Manitoba specimens have also a keel-like appression on outer margin of th 1^1. One specimen referred to *Amplexograptus inuiti* by Jackson (1973: text-fig 2D) has also a mesial spine on th 1^1 in addition to the keel-like structure. This sort of structure has not been observed in topotype specimens of *P. manitoulinensis*, but a mesial spine has been reported and figured by Walters (1977) in specimens from the Lorraine Group of the St Lawrence Lowlands.

The name *Paraclimacograptus decipiens* is proposed below for the short, stubby biserial graptolites which stratigraphically follow on *A. latus* in the upper Vauréal Formation (Fig. 1). *P. decipiens* is morphologically close to *A. prominens* for which it may be easily mistaken (hence its specific name), but its thecae are of the paraclimacograptid type with clearly everted thecal apertures and reduced genicular flanges supported by two short genicular spines (Fig. 2s). The development of the proximal end is of the prosoblastic type and th 1^1 lacks a mesial spine, much as in *A. prominens*. The problem now arises as to the proper generic affiliation of the new species, which could be either in the genus *Paraclimacograptus* Přibyl, 1948 or *Paraorthograptus* Mu, 1974. *Paraclimacograptus* has *P. innotatus* (Nicholson) as type species. *P. innotatus* (Figs 5l–n) is a thin, short graptolite, restricted to the lower Llandovery, with an advanced prosoblastic type of proximal-end development, thecae slightly inclined to the axis of the rhabdosome with wide apertural excavations, everted thecal apertures and short genicular processes which turned out to be flanges in isolated Siberian specimens (Crowther 1981: pl. 13, fig. 4). It lacks a mesial spine on th 1^1. Rickards (1970: 32) has also noted a complete median septum on deformed specimens identified as *C. innotatus*, but it is probably the trace of the virgula. *Paraorthograptus* has *P. typicus* Mu from the Upper Ordovician *Paraorthograptus uniformis* Zone of the Wufeng Shale of central China as type species. This species was described as having '... thecae of the orthograptid type with paired ventral spines ... pointed obliquely downward at the proximal end, horizontal at the distal end ... Interthecal septa straight, slightly inclined, not curved; apertural margins everted, not horizontal ...' (Mu *et al.* 1974: 161; translated). No mention was made of the proximal end, which is not preserved in the holotype specimen (Fig. 5a); it is preserved, however, on a complete specimen on the type slab (Fig. 5b) and shows an apparently advanced type of proximal-end development, much as in *Paraorthograptus pacificus* (Ruedemann) (Figs 5c–f). The type species of *Paraclimacograptus* and *Paraorthogratus* share the same basic rhabdosome morphology, *i.e.* a prosoblastic type of proximal-end development, thecae inclined to the rhabdosome axis and wide thecal excavations with everted apertural margins. They differ, however, in the type and size of genicular processes which are flanges in species assigned to *Paraclimacograptus* (Fig. 5j) and genicular spines of various length in species included into *Paraorthograptus*. The latter also have a mesial spine on th 1^1, a virgella and antivirgellar spines, whereas the former generally lack a mesial spine on th 1^1 (except in some specimens of *P. manitoulinensis* figured by Walters 1977) and also, apparently, antivirgellar spines in the younger species such as *P. innotatus* (see Crowther 1981: pl. 13, fig. 4). The problem is whether two genera are needed to group species on the basis of external morphology, conspicuous as it may be. Lin & Chen (1984: 216), for instance, have tried to solve this problem by simply assigning *Climacograptus innotatus* Nicholson to *Paraorthograptus* in describing Chinese specimens identified and figured as *Paraorthograptus innotatus* (Nicholson). However, a study of the Chinese specimens has revealed that they are fragmentary growth or juvenile stages of *P. typicus*. One of them, complete with mesial spine on th 1^1 and long, paired genicular spines, is shown here as Fig. 5k. This deviation notwithstanding, I feel that the genus *Paraclimacograptus* should group species characterized by a prosoblastic proximal development (advanced as in the type species or more primitive as in *P. manitoulinensis*), thecae inclined to the rhabdosome axis, wide thecal excavations, everted apertures and genicular flanges. The genus *Paraorthograptus* should group all species which, in addition to the basic morphology of the paraclimacograptids, have genicular spines rather than flanges, a mesial spine on th 1^1 and antivirgellar spines. *Paraclimatograptus decipiens* has genicular processes consisting of reduced flanges supported by short, lateral spines (Fig. 2s). It may be regarded as a transitional form between species assigned to *Paraclimacograptus* and *Paraorthograptus*, but the fact that flanges are still present, genicular spines poorly developed and the rhabdosome lacks a mesial spine on th 1^1 support its inclusion in *Paraclimacograptus*, and it will be so described below.

The following graptolites have been identified from the *P. prominens* Zone from surface outcrops and the N.A.C.P. drill core (Fig. 1): *Amplexograptus latus* (Elles & Wood), *Amplexograptus prominens* Barrass, *Paraclimacograptus decipiens* n.sp., *Glyptograptus* cf. *G. hudsoni* Jackson, *Peiragraptus fallax* Strachan, *Rectograptus abbreviatus* (Elles & Wood), *Orthograptus*? and *Desmograptus* sp. In the N.A.C.P. well (Riva 1969), *Amplexograptus latus* has a short, 34 m long range at the base of the *P. prominens* Zone, from the 2047 to the 1933 ft level (614–579 m), whereas *P. decipiens* ranges through the middle and upper part of the zone, from the 1647 to the 1376 ft level (493–412 m), for a total of at least 80 m. *Glyptograptus* cf. *G. hudsoni* (Figs 2k–n) was described by Jackson (1971) from the Upper Ordovician of Southampton Island, north of Labrador and Akpatok Island; in the N.A.C.P. well it has a long range extending through both the *D. complanatus* and the *A. prominens* Zones to terminate somewhere in the upper Vauréal Formation (Fig. 1), for a total of at least 650 m; *P. fallax* is a rare graptolite and has been recognized in only one collection from the mouth of the Patate River in association with *A. latus*, *R. abbreviatus* and *G.* cf. *G. hudsoni* (Riva & Petryk 1981); *R. abbreviatus* occurs sporadically through both the *D. complanatus* and *A. prominens* Zones and two specimens were also collected by A. A. Petryk from member 6 of the Ellis Bay Formation, just below the Ordovician–Silurian boundary (Fig. 3i).

Correlation of the *A. prominens* Zone. *A. latus* is a cosmopolitan graptolite long recorded from the *D. anceps* Zone of southern Scotland and, especially, the *D. complexus* and *P. pacificus* Subzones (Williams 1982). This allows us definitely to correlate the *A. prominens* Zone of Anticosti Island with the uppermost British Ordovician. *A. latus* also occurs in the *C. supernus* Zone of Kazakhstan (Koren *et al.* 1980), where it has been described as *Amplexograptus stukalinae*, the *C. pacificus* Subzone of the Omulev Mountains of Siberia (Koren *et al.* 1983), where it is represented by *A. latus hekandaensis*, the *Amplexograptus yangtzensis* to the *Diplograptus bohemicus* Zones of the Wufeng Shale of central China (Mu & Lin 1984), where *A. latus* has been called *A. suni* and *A. yangtzensis* (Fig. 4a), and from the Bolindian *D. ornatus* and *C. latus* Zones of Victoria, Australia (VandenBerg 1981*a*). The *A. prominens* Zone of Anticosti is correlated with all the above-mentioned zonal levels (Fig. 1).

Graptolites from the Ellis Bay and lower Becscie Formations. Graptolites are scarce above the *A. prominens* Zone. Few graptolites have been collected above member 2 of the Vauréal Formation besides a few specimens of *G.* cf. *hudsoni* (Figs 2l–n). Graptolites are also scarce in the Ellis Bay and Becscie Formations: the few collected are either indicative of the uppermost Ordovician or are long-ranging species that straddle the Ordovician–Silurian boundary. Members 4 and 7 of the Ellis Bay Formation have yielded fragmental climacograptids which I have assigned to *Scalarigraptus angustus* (Elles & Wood); one of them is shown as Fig. 3i. In Scotland this graptolite ranges through the *D. anceps* Zone (Williams 1983: fig. 3) and may be taken to indicate that member 6 of the Ellis Bay is of uppermost Ordovician age. At Salmon River, the Becscie Formation has yielded fragments of *S. angustus* from its contact with the top of reef structures of the Ellis Bay upwards (Fig. 1). An excellent three-dimensional specimen of *S. angustus* was collected by A. A. Petryk a few metres above the base of the Becscie (Figs 3t, u); two small collections of this species were made 7 and 30 m above the base of the formation at pool 9 on Salmon River (Figs 3j–s) and one specimen (Fig 3v) was collected from the Gun River Formation, well above the Ordovician–Silurian boundary. This is the longest specimen of *S. angustus* collected on Anticosti Island. *S. angustus* ranges from the Ashgill to the lower Llandovery, and it is common in the *D. anceps*, *G. persculptus*, *A. acuminatus* and other Zones at or above the Ordovician–Silurian boundary and cannot be regarded as a good zonal indicator. In closing, it will be noted that a large climacograptid approaching *Scalarigraptus normalis* (Lapworth) in size (Fig. 3w) was collected by T. E. Bolton from the basal Becscie Formation on the east side of Ellis Bay at Cap-à-l'Aigle. *S. normalis* is only known from the *G. persculptus* Zone to the lower Llandovery.

The new genus *Scalarigraptus*. The occurrence of graptolites of the *normalis* (or *scalaris*) group in the Ellis Bay and Becscie Formations brings again to the fore the problem of their generic affiliation which cannot any longer be the traditional polyphyletic genus *Climacograptus* Hall. *Climacograptus* was created by Hall (1865: 111–112) for 'simple stipes with sub-parallel margins having a range of cellules (thecae) on each side', which were to be 'short and square'. *Graptolithus bicornis* was designated as the type species and the members of the *G. scalaris* group of Linné were 'conceived' as the 'veritable species of this genus'. (The generic name *Climacograptus* was obtained by adopting the Greek noun *klimax*, equivalent to the Latin *scala*, ladder, of which *scalaris* is the adjective). Since its creation, this genus has known enormous popularity, having been used as a generic umbrella for all sorts of biserial graptolites characterized by square or climacograptid thecae, at least in the mature or distal part of the rhabdosome. Elles & Wood (1906) attempted to deal with the large number of British graptolites assigned to *Climacograptus* by dividing them into five groups on the basic of thecal outline, type of apertural excavation or thecal ornaments such as spines, but did not propose new genera or subgenera. Přibyl (1947, 1948), on the other hand, went a step further and proposed the genus *Pseudoclimacograptus* for climacograptids characterized by a zig-zag median septum connected by transverse rods to the thecal septa and the genus *Paraclimacograptus* for climacograptids with genicular spines. The genus *Pseudoclimacograptus* has since been widely accepted by graptolite specialists, but the genus *Paraclimacograptus* has been overshadowed by the genus *Paraorthograptus* Mu, 1974. Riva (1974*b*, 1976) showed, on the basis of three-dimensional topotype material, that *C. bicornis* had a primitive diplograptid, or streptoblastic, type of proximal-end development and thus differs significantly from other climacograptids with a prosoblastic type of proximal-end development. The graptolites of the *scalaris* group, considered by Hall (1865: 112) as the 'veritable species' of *Climacograptus*, have an advanced prosoblastic type of proximal-end development and cannot be regarded as true climacograptids, although they share with *C. bicornis* a similar distal development. For this reason, I am proposing the new genus *Scalarigraptus* for all graptolites of the '*scalaris*' or *normalis* group and for all Ordovician climacograptids with an advanced prosoblastic type of proximal-end development and with a septate or partly septate rhabdosome. *C. normalis* will be designated as the type species of the new genus.

In 1949 Obut erected the genus *Hedrograptus* for early Silurian climacograptids with insignificant or incomplete apertural excavations on one side of the rhabdosome and complete on the other. The figures of the type species, *H. janischewskyi* Obut, show that it is also characterized by an advanced prosoblastic type of proximal-end development, just like *C. normalis.* In 1975, Obut extended *Hedrograptus* to include all climacograptids of the *scalaris* group. This would mean that *Hedrograptus* rather than the proposed *Scalarigraptus* is actually the genus intended for climacograptids of the *scalaris* group. I have, however, thanks to the cooperation of A. M. Obut, been able to examine a latex cast of the holotype of *H. janischewskyi* (Fig. 6a) and conclude that *Hedrograptus* is based on an incomplete and distorted specimen preserved in $\frac{3}{4}$-face view which does not allow us to ascertain whether the thecae are climacograptid or glyptograptid. Another specimen from the type locality, also preserved in $\frac{3}{4}$-face view (Fig. 6b), is much larger than the holotype of *H. janischewskyi* and probably not conspecific with it. For these reasons, I have been reluctant to adopt *Hedrogratus* and propose instead the genus *Scalarigraptus.*

Systematic palaeontology

Family **DIPLOGRAPTIDAE** Lapworth, 1873

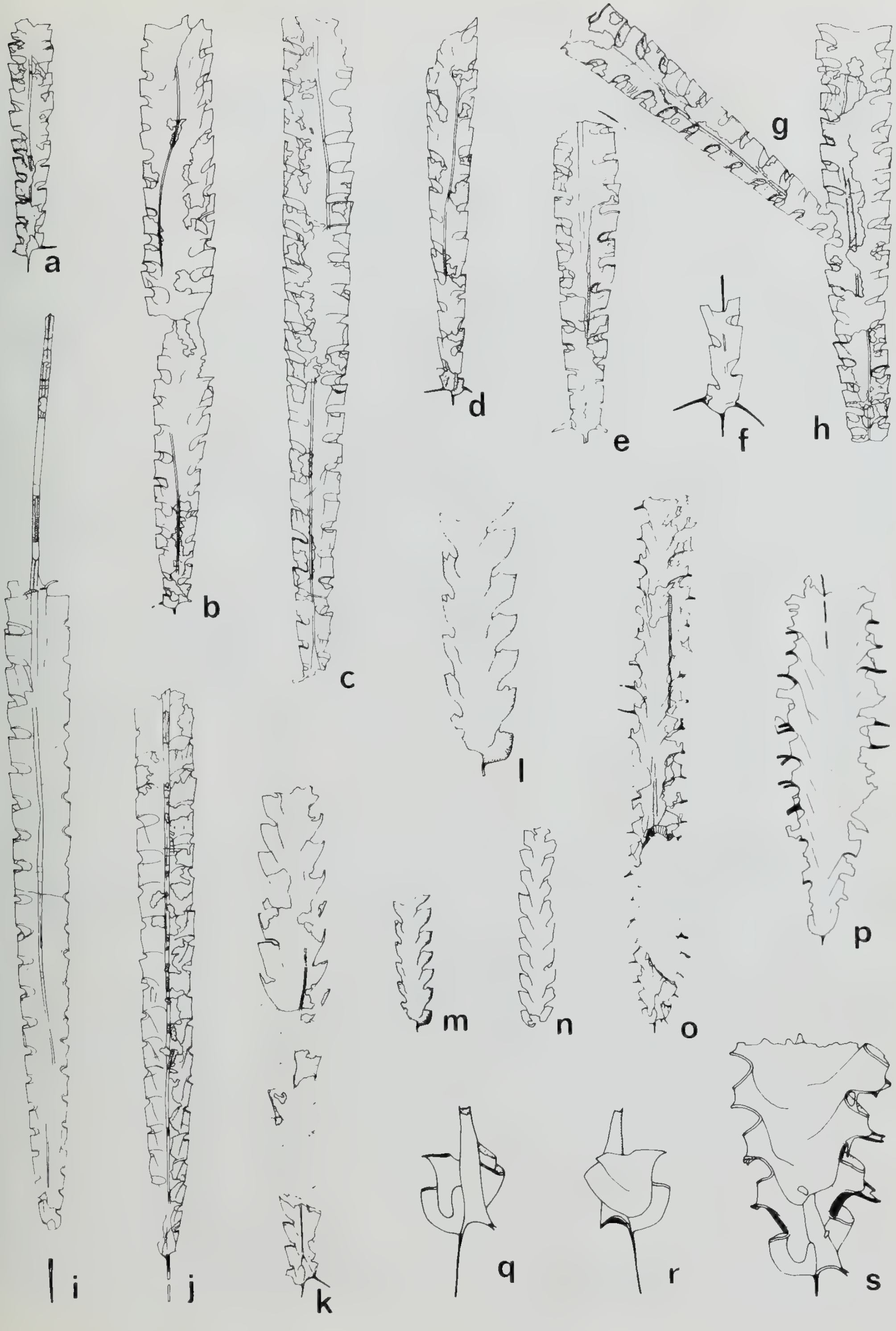
Genus ***AMPLEXOGRAPTUS*** Elles & Wood, 1907

Amplexograptus latus (Elles & Wood)

Figs 2a–h, 4

1906 *Climacograptus latus* Elles & Wood: 204–205; pl. 27, figs 3a–e and g, *non* figs 3f–h; text-figs 135a–d.
1933 *Climacograptus inuiti* Cox: 1–19, pls 1, 2.
1953 *Amplexograptus elongatus* Barrass: 62–66; figs 6–8.
non 1970 *Climacograptus latus* Elles & Wood; Toghill: 22; pl. 15, figs 1, 2.
1974 *Amplexograptus disjunctus yangtzensis* Mu & Lin; Mu *et al.*: 162; pl. 70, fig. 6.
1980 *Amplexograptus stukalinae* Mikhailova; Koren *et al.*: 125–126; pl. 4, figs 1, 2.
1982 *Climacograptus latus* Elles & Wood; Williams: 39–40; pl. 3, figs 12–18. [See also for other synonyms.]
1983 *Climacograptus latus hekandaensis* subsp. nov.; Koren & Sobolevskaya: 116–117; pl. 30, figs 2–6; pl. 31, figs 1–3.
1983 *Climacograptus latus* Elles & Wood; Wang *et al.*: pl. 3, fig. 1.
1984 *Amplexograptus suni* (Mu); Mu & Lin: 56; pl. 5, figs 4–6.

Fig. 2 Type specimens of *Amplexograptus latus* (Elles & Wood, 1906) and graptolites from the Vauréal Formation. a–h, Type specimens of *A. latus* from the upper Hartfell Shale, Main Cliff, Dob's Linn; a, SM 19683b (Elles & Wood 1906: text-fig. 135a), paralectotype, × 5; b, SM A19680 (Elles & Wood 1906: pl. 27, fig. 3a), proposed **lectotype**, × 5; c, BU 1195 (Elles & Wood 1906: pl. 27, fig. 36), paralectotype, × 5; d, SM 19682a (Elles & Wood 1906: pl. 27, fig. 3g, text-fig. 135c), paralectotype, × 5; e, BU 1412b, unfigured paralectotype (on the same slab as BU 1412a of Fig. 2j), × 5; f, SM A19683c, unfigured growth stage, × 10; g, BU 1411a (Elles & Wood: pl. 27, fig. 3e), paralectotype, × 5; h, BU 1411b, unfigured paralectotype, × 5; i–j, *Scalarigraptus tubuliferus* (Lapworth) originally included in the type material of *A. latus*; i, BU 1413 (Elles & Wood: pl. 27, fig. 3h) doubtfully included, × 5; j, BU 1412a (Elles & Wood 1906: pl. 27, fig. 3f), × 5; k–n, *Glyptograptus* cf. *G. hudsoni* Jackson; k, G.S.C. 82880, from the 2739 ft (822 m) level in the N.A.C.P. core, × 5; l, m, G.S.C. 82881, from member 2 of the Vauréal Formation at Cap Crotté, Anticosti Island (A. A. Petryk's coll. 76 AP29-1), respectively × 10 and × 5; n, G.S.C. 82882, same locality and collection, × 5; o–s, *Paraclimacograptus decipiens* sp. nov.; o, G.S.C. 82883, **holotype**, longest specimen recovered from the 1376 ft (413 m) level in the N.A.C.P. core × 5; p, G.S.C. 82884, paratype, a large macerated specimen (A. A. Petryk's coll. 83 AP6-5), from 90 m above the mouth of Patate River, member 2, Vauréal Formation, × 5; q–s, G.S.C. 82885, 82886, paratypes, isolated specimens from the 1381 ft (414 m) level in the N.A.C.P. core showing the development of the proximal-end thecal structure, × 15. Note the development of vertical cortex filaments in the apertural excavations of th 2^2 and 3^2.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LECTOTYPE. SM A19680 (Fig. 2b) (Elles & Wood 1906: pl. 27, fig. 3a) from the upper Hartfell Shale, *D. anceps* Zone, Main Cliff, Dob's Linn, Scotland. Herein selected.

PARALECTOTYPES. SM A19683b and A19682a (Figs 2a, b), BU 1195 and 1411a (Figs 2c, g), BU 1414 and 1196 (not figured because of poor preservation) and the following specimens from the type collection, not previously figured: BU 1412b (Fig. 2e), 1411b (Fig. 2h) and a growth stage, SM A19683c (Fig. 2f). BU 1413 and 1412 (Figs 2i, j) have been excluded from *A. latus* and assigned to *C. tubuliferus*.

OTHER MATERIAL EXAMINED. Several topotype specimens of *A. inuiti* from Akpatok Island, the N.A.C.P. drill core and surface collections made by A. A. Petryk from member 2 of the Vauréal Formation, Anticosti Island. The type and topotype material of *Amplexograptus stukalinae* Mikhailova and of *Climacograptus latus hekandaensis* Koren & Sobolevskaya stored either at the VSEGEI in Leningrad or at the Institute of Geology and Palaeontology of the Akademya Nauk, Moscow, U.S.S.R.; the type or topotype material of *Amplexograptus suni* (Mu) and *Amplexograptus disjunctus yangtzensis* Mu & Lin at the Institute of Geology and Palaeontology, Academia Sinica, Nanjing, and at the Institute of Geology and Mineral Resources, Academy of Geological Sciences, Yichang, China.

DESCRIPTION. Rhabdosome up to 5 to 6 cm in length, gradually widening from 0·8–1·1 mm at the level of th 1^2 aperture to a maximum of 2·2–2·4 (exceptionally 2·6) mm distally, attained within 2 or 3 cm. The average width, however, is less than 2 mm, generally 1·6–1·8 mm. A waist-like constriction may also be noted in some specimens above the the first pair of thecae. Thecae 14–12 in 10 mm proximally, decreasing to 11–12 distally. Development of proximal end of prosoblastic type (Cox 1933: 6, 7; figs 1–21). The sicula is 1·5 mm long; it secretes a virgella and two antivirgellar spines. Th 1^1 originates low in the metasicula, grows down along the virgellar side to the sicular aperture, then turns out and upwards, secreting a mesial spine at the point of upward growth; th 1^2 buds off from the downward-growing portion of th 1^1, grows around the reverse side of the sicula to turn up at the point of issuance of the antivirgellar spines (Fig. 4). Th 2^1 buds off th 1^1 and th 2^2 from th 1^2 and so on alternately to the distal end of the aseptate rhabdosome. Thecae are of the amplexogratid type with apertural lappets and thecal excavations occupying about $\frac{1}{4}$ of the rhabdosome width. A selvage runs around the thecal apertures and the infragenicular walls to form a short genicular flange.

REMARKS. The type material of *A. latus* was mixed, containing two specimens herein assigned to *C. tubuliferus* (Figs 2i, j). Because of its world-wide distribution, this species has been identified and described as *C. latus* and also under a number of names such as *C. inuiti* and *A. stukalinae* Mikhailova, *Climacograptus latus hekandaensis* Koren & Sobolevskaya for specimens from Kazakhstan and NE Siberia, and as *Amplexograptus disjunctus* Mu & Zhang, *Climacograptus suni* (Mu) and *Amplexograptus disjunctus yangtzensis* Mu & Lin for specimens from the Upper Ordovician Wufeng Shale of central China. *A. yangtzensis* is a species in its own right and not a subspecies of *A. disjunctus*, a *nomen nudum*, the type of which could not be located in a recent study visit to Nanjing. It is based on a single three-dimensional specimen (Mu *et al.* 1974: pl. 70, fig. 4), here refigured as Fig. 4a, from a zone of the same name in the lower Wufeng Shale, where graptolites are generally preserved in relief in a black shale. Farther up in the Wufeng Shale, *A. yangtzensis* is replaced by *A. suni*, which differs from *A. yangtzensis* only in being preserved as flattened, brown, flaky films.

The specimens from Dob's Linn, Scotland, identified as *C. latus* by Toghill (1970) belong to either *S. normalis* or *S. tubuliferus*.

STRATIGRAPHICAL AND GEOGRAPHICAL OCCURRENCE. *A. latus* is restricted to the *D. anceps* Zone of Scotland (Williams 1982) and may be considered as one of its diagnostic fossils. It is a widely distributed, cosmopolitan species, known from the *C. supernus* Zone of Kazakhstan and NE Siberia, the Upper Ordovician of China and correlative strata elsewhere. In SE Australia it helps name the upper Bolindian *D. ornatus*–*C. latus* Zone (VandenBerg 1981*a*).

Genus *PARACLIMACOGRAPTUS* Přibyl, 1948

TYPE SPECIES (by original designation). *Climacograptus innotatus* Nicholson (Nicholson 1869: 238; pl. 11, figs 16, 17).

DIAGNOSIS (amended from Přibyl 1948: 40–41, 47–48, fig. 6). Rhabdsome aseptate, apparently ovoid in cross-section; thecae of the paraclimacograptid type, inclined to the axis of the rhabdosome; apertural excavations wide and deep with everted thecal apertures and genicular flanges, strengthened by a selvage (list) split into two short spines at the geniculum in some species. Proximal end characterized by a prosoblastic type of development, and provided with a virgella, antivirgellar spines and, exceptionally, a mesial spine on th 1^1 (in older species).

INCLUDED SPECIES. The following species may be included in *Paraclimacograptus: Paraclimacograptus innotatus* (Nicholson), *Paraclimacograptus manitoulinensis* (Caley), *Paraclimacograptus decipiens* sp. nov., *Paraclimacograptus* sp., an undescribed species from the *Climacograptus wilsoni* Zone of Gaspé, Canada.

Climacograptus innotatus nevadensis Carter (Riva 1974*a*: figs 2k–m) from the late mid-Ordovician of Nevada, Texas (Marathon region), Oklahoma (unpubl. data) and Australia (VandenBerg 1981*b*) is close to *Scalarigraptus.* This species has an advanced prosoblastic proximal-end development, thecae of the climacograptid type with stiff genicular spines in the first six to twelve pairs, a long virgella accompanied by a sicular downgrowth, a long inflated virgula and a sicula lacking the prosicula. These characteristics brings it closer to the scalarigraptids of the *tubuliferus* group of the Upper Ordovician rather than to the paraclimacograptids.

Paraclimacograptus decipiens sp. nov.

Figs 2o–s

HOLOTYPE. G.S.C. 82883 (Fig. 2o), from the 1376 ft (413 m) level in the N.A.C.P. core, upper Vauréal Formation, Anticosti Island.

PARATYPES. G.S.C. 82884 (Fig. 2p), from 90 m above the mouth of Patate River, Anticosti Island, member 2 of the Vauréal Formation; G.S.C. 82885 and 82886 (Figs 2q–s), isolated growth stages from the 1381 ft (414 m) level of the N.A.C.P. core, upper Vauréal Formation.

NAME. Latin *decipiens*, deceiving.

DESCRIPTION. Rhabdosome of moderate length, usually not exceeding 2 to 3 cm, maximum observed 4 cm (Fig. 2o), widening rapidly from 0·8–1·0 mm at the level of the aperture of th 1^2 to 1·6–2·0 mm (maximum observed 2·4 mm) at the level of the 4th to 5th pair of thecae. Thecae numbering 8 in 5 mm, or 15 in 10 mm, proximally, decreasing to 12–13 in 10 mm distally, of the paraclimacograptid type with everted thecal apertures, except for the first two which have low lappets (faintly visible also on the second pair of thecae in Fig. 2s). Interthecal septa inclined at 20° to 40° to the rhabdosome axis; supragenicular walls parallel or slightly inclined to it. Thecal excavations wide, occupying $\frac{1}{4}$ of the rhabdosome width, reinforced by a selvage running around the thecal aperture and the infragenicular wall and terminating as two short, stiff genicular spines supporting a reduced hood (Fig. 2s). Development of the proximal end of the prosoblastic type. Sicula about 1·5 mm long, partly exposed on the obverse side of the rhabdosome (Figs 2o, s). Th 1^1 originates low in the metasicula, grows down the virgellar side to the sicular aperture before turning out and upwards to terminate about level with its point of origin. Th 1^2 buds off the downward-growing portion of th 1^1, grows diagonally around and up on the obverse side of the rhabodosome (Fig. 2r); th 2^1 buds off from th 1^2 and th 2^2 from th 1^2 and so on alternately to the distal end of the rhabdosome. A thin nema passes through the rhabdosome and extends a short distance beyond it. The rhabdosome is aseptate.

REMARKS. The development of the proximal end of *P. decipiens* is identical to that of *A. latus* and *A. prominens*, suggesting a close genetic relationship between the three species. *P. decipiens* is much larger than *P. innotatus* which has a proximal development of the advanced prosoblastic type. *P. decipiens* is much closer to *P. manitoulinensis* from the lower Upper Ordovician of NE North America (Riva 1969) (Figs 5g, h and i), but this species is thinner, of uniform width and has genicular flanges strengthened by a thickened selvage (Fig. 5j). A mesial spine on th 1^1 may occur sporadically in some rhabdosomes (Walters 1977).

STRATIGRAPHICAL AND GEOGRAPHICAL OCCURRENCE. *P. decipiens* is only known from the *A. prominens* Zone of Anticosti, where it has a stratigraphical range of at least 80 m in the upper Vauréal Formation (Riva 1969). It has been found also sporadically in recent surface collections made by A. A. Petryk and in an older collection (Y.P.M. 3036/4) made by W. H. Twenhofel and stored at the Peabody Museum of Yale University (Riva & Petryk 1981: 160).

Genus ***SCALARIGRAPTUS*** nov.

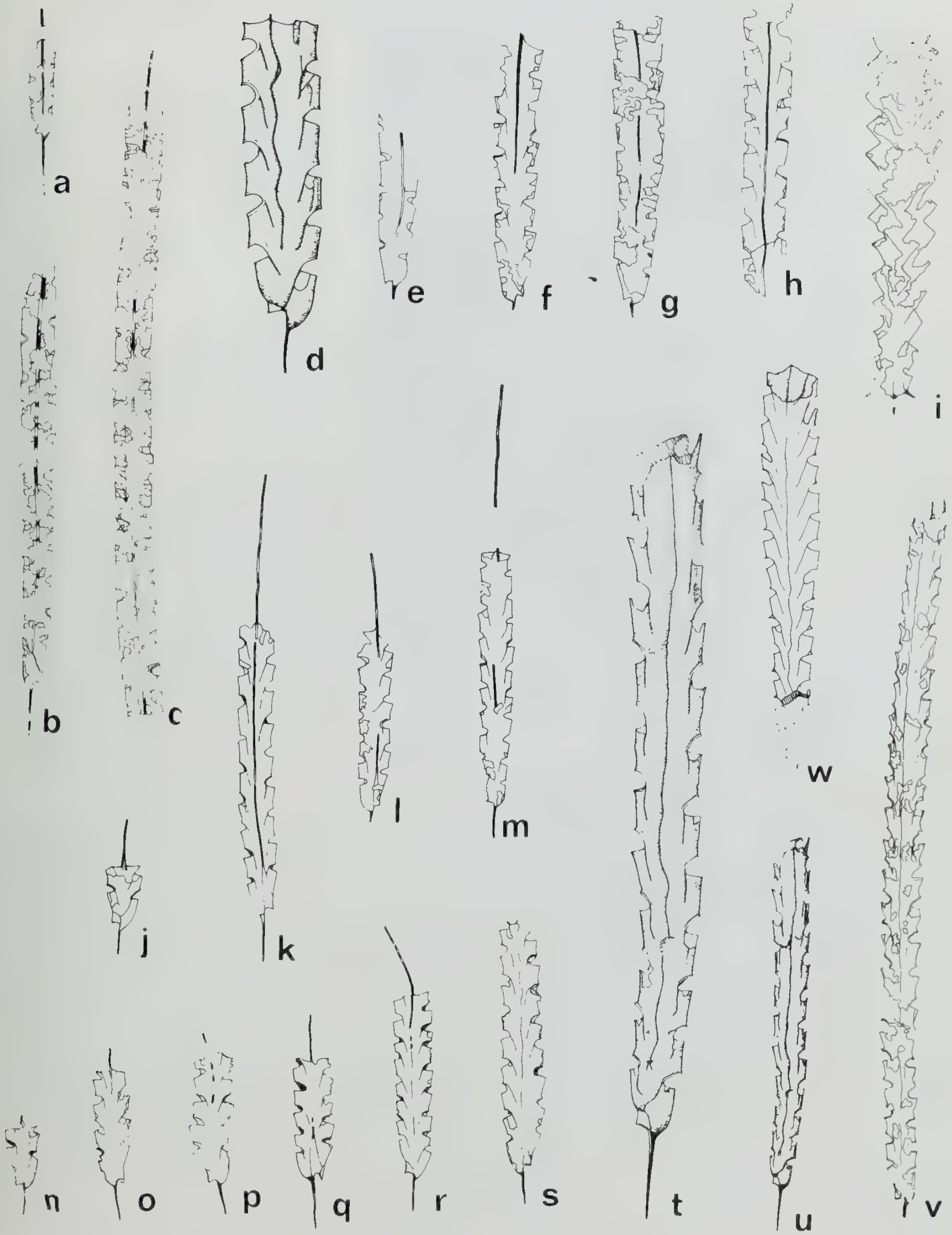
TYPE SPECIES. *Climacograptus normalis* Lapworth (Lapworth 1877: 138; pl. 6, fig. 31; Elles & Wood 1906: pl. 26, fig. 2a; Williams 1983: text-fig. 4a).

NAME. From the Latin *scalaris*, ladder-like.

DIAGNOSIS. Rhabdosome septate or partly septate, ovoid to subrectangular in cross-section; thecae of the climacograptid type with definite genicula, deep horizontal apertural excavations and straight supragenicular walls, usually parallel to the axis of the rhabdosome. Proximal-end development of the advanced prosoblastic type with only th 1^1 initially growing down along the sicula. The virgella is the only proximal spine.

INCLUDED SPECIES. The following species, among others, fall within the limits of the diagnosis of *Scalarigraptus*: *C. normalis*, *C. angustus* (Perner), *C. transgrediens* Waern, *C. medius* Törnquist, *C. praemedius* Waern, *C. rectangularis* M'Coy, *C. brevis* Elles & Wood, *C. putillus* (Hall), *C. tubuliferus* Lapworth, *C. nevadensis* Carter, *C. yumenensis* Mu and *C. biformis* (Mu & Lee).

Fig. 3 Syntypes of *Climacograptus miserabilis* Elles & Wood, 1906 and graptolites from the Ellis Bay and the lower Becscie Formations. a–c, Syntypes of *C. miserabilis*; a, BU 1148b (Elles & Wood 1906: text-fig. 120b), proximal end with long virgella (freed from matrix), × 5; b, BU 1150 (Elles & Wood 1906: text-fig. 120a), typical specimen with long virgella (freed from matrix), × 5; c, BU 1146a (Elles & Wood 1906: pl. 26, fig. 3b and text-fig. 120c), distal fragment showing thread-like virgula passing through the thin rhabdosome, × 5. d–h, *Scalarigraptus angustus* (Perner) from the Ellis Bay Formation; d, G.S.C. 82887, obverse view of growth stage preserved in relief, showing climacograptid thecae and wavy median septum, from the oncolite platform bed, basal member 7 (A. A. Petryk's collection 84AP8-2-1F), Pointe Laframboise, Cape Henry, × 10; e–h, G.S.C. 82888–82891, large distorted or fragmentary rhabdosomes from upper member 4 (A. A. Petryk's collection 81AP3-2), Baie des Navots, Ellis Bay, × 5. i, G.S.C. 82892, *Rectograptus abbreviatus* (Elles & Wood), macerated specimen from member 5, Ellis Bay Formation, immediately below reef bioherms, 7 km upriver from mouth of Salmon River, right bank (A. A. Petryk's collection 75APt3-3), × 5. j–m, *S. angustus* (Perner) from the basal beds of the Becscie Formation; j, k, G.S.C. 82893, 82894, a growth stage and an adult individual showing a thin virgella distally prolonged (A. A. Petryk's collection 81AP13-1-1F), from pool 9, Salmon River, 13 m above the base of the formation, × 5; l–m, G.S.C. 82895, 82896, from the basal Becscie at pool 9 on Salmon River (collected by J. Riva 1981), × 5. n–s, G.S.C. 82897–82902, growth series of *S. angustus* (A. A. Petryk's collection 79AP48-4), 7 m above base of the Becscie, base of pool 9 on Salmon River, × 5. t, u, G.S.C. 82903, observe view of *S. angustus* preserved in excellent relief, showing wavy median septum in proximal part of rhabdosome (A. A. Petryk's collection 76AP22-30-6′), 2–3 m above base of Becscie Formation on Salmon River, respectively × 10 and × 5. v, G.S.C. 82904, longest specimen of *S. angustus* recovered from the mid-part of the Gun River Formation, 3·5 km from mouth of Chute Creek, eastern Anticosti (A. A. Petryk's collection 75MPt18-L8C-1F), × 5. w, G.S.C. 69157, *Scalarigraptus normalis* (Lapworth), collected by T. E. Bolton in 1981 from the basal Becscie Formation on the east shore of Ellis Bay near Cap-à-l'Aigle, Anticosti Island, × 5.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Scalarigraptus angustus (Perner, 1895)

Figs 3a–u

1895 *Diplograptus* (*Glyptograptus*) *euglyphus* Lapworth var. *angustus* Perner: 48; pl. 8, figs 14a, b.
1906 *Climacograptus scalaris* (Hisinger) var. *miserabilis* Elles & Wood: 186; pl. 26, figs 3a, b, d, e, g, h, *non* figs 3c, f; text-figs a–c.
1951 *Climacograptus angustus* (Perner) Přibyl: 7; pl. 2, figs 2–9.
1975 *Climacograptus angustus* (Perner); Bjerreskov: 23; fig. 9A.
1980 *Climacograptus angustus* (Perner); Koren *et al.*: 131; pl. 37, figs 2–7; text-figs 34a–e.
1983 *Climacograptus angustus* (Perner); Koren & Sobolevskaya: 106–108; pl. 27, figs 1–5; text-fig. 34.
?1983 *Climacograptus mirnyensis* (Obut & Sobolevskaya); Koren & Sobolevskaya: 132–133; pl. 37, figs 2–5; text-figs 47K–H.
1983 *Climacograptus miserabilis* Elles & Wood; Williams: 615–616; text-figs 3f–i, ?j, 4f–i, 5a–b. [See also for a more extended pre-1983 synonymy.]

HOLOTYPE. National Museum of Prague CD 1835, partly figured by Perner (1895: pl. 8, figs 14a–b) and refigured in full by Přibyl (1951: pl. 2, fig. 8).

MATERIAL STUDIED. The type collection of *C. miserabilis* in the Lapworth collection of Birmingham; part of the collections made by P. Toghill at Dob's Linn; the type and topotype material of *C. angustus* in Prague; the collections of *C. angustus* and *C. mirnyensis* at VSEGEI, Leningrad, several collections made by A. A. Petryk from the Ellis Bay, Becscie and Gun River Formations of Anticosti Island.

DESCRIPTION. Rhabdosome up to 2 cm in length, widening imperceptibly from 0·8–0·9 mm at the level of th 1^2 aperture to a maximum of 1·0–1·1 mm (exceptionally 1·2 mm) within one pair of thecae. Thecae of the climacograptid type, numbering 11–12 in the first 10 mm, decreasing to 10–11 distally, with sharp genicula and supragenicular walls parallel to slightly inclined to the rhabdosome axis. Thecal apertures horizontal to slightly everted; thecal excavations wide and semicircular, occupying about $\frac{1}{4}$ of the rhabdosome width and reinforced by a thin selvage around the aperture and the infragenicular walls, terminating as a slight genicular flange (Figs 3d, t). Development of the proximal end of the advanced prosoblastic type. Sicula from 1·2 to 1·6 mm long, secreting a long virgella; it is mostly exposed on the obverse side of the rhabdosome (Williams 1983: text-fig. 3h). Th 1^1 first grows down along the sicula and then turns out and upwards at the sicular aperture (Figs 3d, t); th 1^2 grows up from th 1^1 and th 2^1 from th 1^2. Th 2^1 is also the dycalical thecae which gives rise to two independent linear series separated by a median septum. The median septum begins on the obverse side of the rhabdosome at about the level of th 1^2 aperture (its point of origin is marked by a notch in some specimens, Fig. 3t) and follows a wavy pattern through the first 5 or 6 pairs of thecae before straightening out (Figs 3d and t). A thin, thread-like nema passes through the rhabdosome and extends for some distance beyond.

REMARKS. In 1951 Přibyl pointed out that *C. miserabilis* Elles & Wood 1906 was identical to, and the junior synonym of, *C. angustus* (Perner 1895). This synonymy was accepted by some workers (for instance Bjerreskov 1975: 23) but not by British workers for a number of reasons best summarized by Williams (1983: 616). Recently, I have been able to study the type material of both *C. miserabilis* and of *S. angustus*. *C. miserabilis* is based on seven specimens from the *D. complanatus* Zone and two from the *D. anceps* Zone of Dob's Linn, Scotland. The two specimens from the *D. anceps* Zone do not belong to *C. miserabilis*: one, BU 1145b (Elles & Wood 1906: pl. 26, fig. 3c), is a distal fragment of *tubuliferus*, and the other, BU 1149 (Elles & Wood: pl. 26, fig. 3f), is of uncertain affiliation. The specimens from the *D. complanatus* Zone (three of which are shown here as Figs 3a–c) are preserved as thin, flaky, abraded films. They all belong to *C. miserabilis*. They are from 0·8 to 1·1 mm wide and have 12–11 thecae per 10 mm proximally and 11 distally. The proximal end bears a long virgella, and a thin nema passes through the rhabdosome. This is all that can be learned from the type material of *C. miserabilis*. The type and topotype material of *S. angustus* is more diversified and contains several specimens in partial relief. (I was unable to draw any specimens, but was assisted in my work by Dr A.

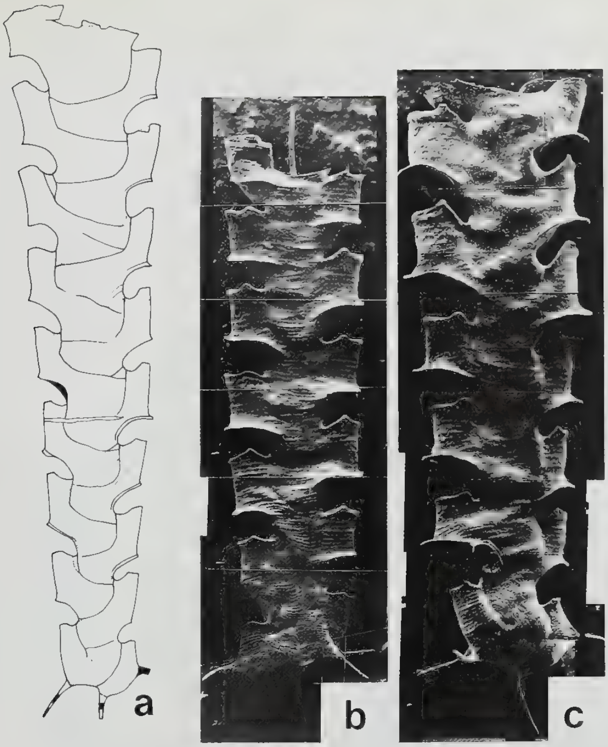

Fig. 4 a, N.I.G.P. Catalogue Number 21410, holotype of *Amplexograptus yangtzensis* Mu & Lin (=*A. latus*), × 20; b and c, SEM montages of *Amplexograptus inuiti* (Cox) (=*A. latus*) from Akpatok Island, Canada: b, SM A102524, obverse view; c, SM A102521, reverse view, both ×20 (courtesy of Peter Crowther).

Přibyl). The specimens attain a width of 1·0–1·1 mm, have 12–11 thecae per 10 mm proximally and 10 distally. The thecae are all of the climacograptid type with strong genicula. The proximal end bears a long virgella and a thin virgula passes through the rhabdosome. The holotype is a complete, not partial, specimen as claimed by Strachan (1971: 34); it has been refigured in full by Přibyl (1951: pl. 2, fig. 8). With the aforesaid in mind, I do not see any morphological differences between the types of *C. miserabilis* and *S. angustus* and do not hesitate to place the former in synonymy with the latter.

The specimens from the basal Becscie Formation (Figs 3j–u) are all practically identical to the type of *S. angustus* and so are those from the Gun River Formation. The specimens from member 4 of the Ellis Bay Formation (Figs 3e–h) are wider (from 1·1 to 1·3 mm) because of poor preservation and distortion; that from the base of member 7 (Fig. 3d) has the same dimensions as the holotype in Prague.

Stratigraphical and geographical occurrence. *S. angustus* is a cosmopolitan graptolite ranging through the Upper Ordovician and part of the Lower Silurian. In NE Siberia (Omulev Mountains) it is common from the base of the *C. extraordinarius* Zone to the top of the *A. acuminatus* Zone (Koren *et al.* 1983: figs 62, 64). On Anticosti Island it first occurs at the top of the *P. manitoulinensis* Zone (Riva 1969: figs 11, 13), below the base of the *D. complanatus* Zone, and extends all the way up into the Gun River Formation of mid-Llandovery age.

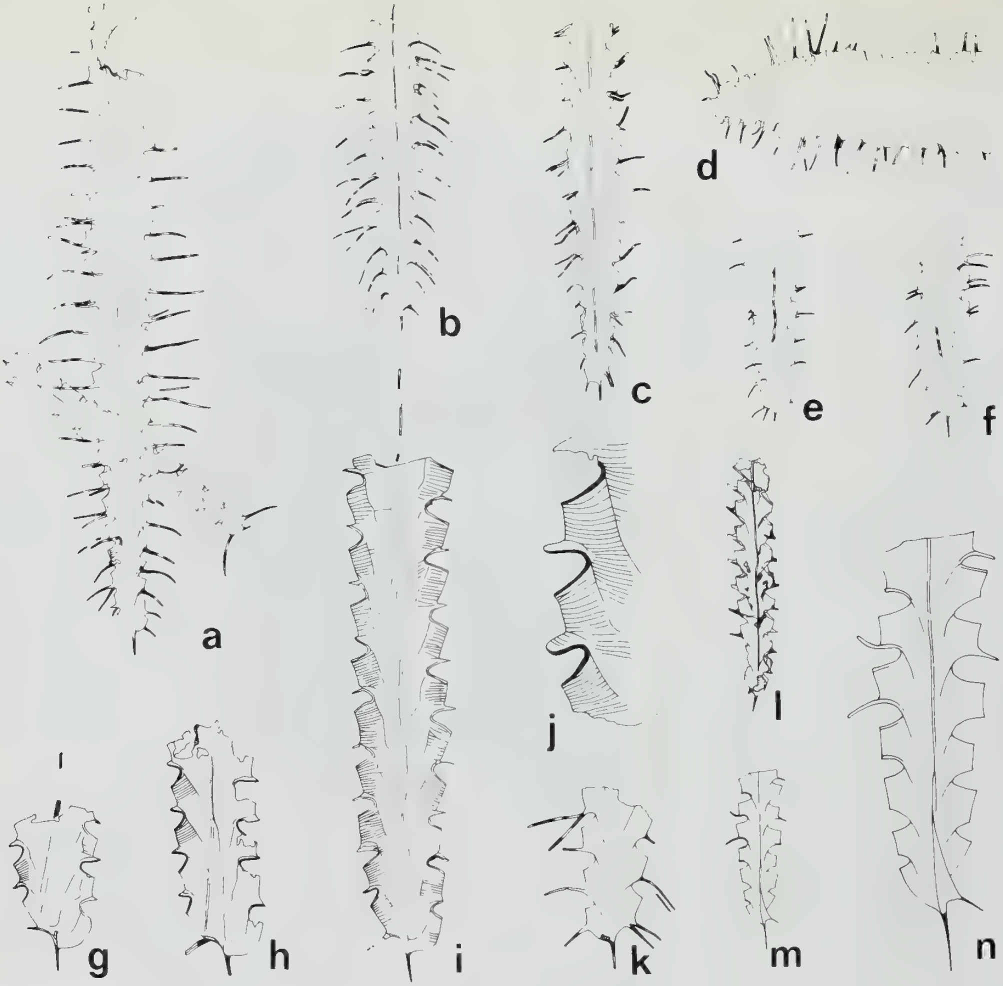

Fig. 5 a, b, *Paraorthograptus typicus* Mu; a, N.I.G.P. Cat. No. 21418a, counterpart of the holotype (better preserved than the part) from the Wufeng Shale north of Yichang, central China, showing the characteristic long, paired genicular spines of the species but with the proximal end missing (a rhabdosome of *Climacograptus longispinus supernus* Elles & Wood lies diagonally across its proximal end), × 5; b, unfigured specimen of *P. typicus*, with a complete proximal end, occurring on the same slab as the holotype, × 5. c–f, U.S.N.M. 415038–415401, rhabdosomes of *Paraorthograptus pacificus* (Ruedemann) from the Phi Kappa Formation at Trail Creek, Idaho, U.S.A., near the type locality of the species, showing their characteristic short genicular spines, both paired and triple, and stubby form; note the tectonic deformation undergone by specimens of Figs 5c and d lying normal to each other, × 5. g–j, G.S.C. 56899, 56895, 56900 and 56901, respectively, topotypes of *Pseudoclimacograptus manitoulinensis* (Caley) from the upper Whitby Formation, 5 km south of Little Current west side of Rt 68, Manitoulin Island, Ontario, Canada; g–i, growth series showing distinct fusellar rings, × 10; j, detail of thecal excavations showing everted thecal apertures and well-developed genicular lappets strengthened by a selvage, × 20. k, N.I.G.P. Cat. No. 82816, proximal end of *P. typicus* figured as *Paraorthograptus innotatus* (Nicholson) by Lin & Chen (1984: pl. 4, fig. 7), showing the spinose processes typical of the species: virgella, antivirgellar spines, mesial spine on th 1^1 and genicular spines, × 10. l–n, *Paraclimacograptus innotatus* (Nicholson), topotypes from the lower Birkhill Shale (Lower Silurian) at Dob's Linn, southern Scotland; l, SM A20222, specimen figured by Elles & Wood (1906: pl. 27, fig. 10a) as a 'typical specimen' (but not the 'type' of Nicholson), × 5; m, n, SM A20232 (*op. cit.*: pl. 27, fig. 106), specimen showing advanced prosoblastic development of proximal end and a partly uncovered sicula below th 1^2, × 5 and × 10, respectively.

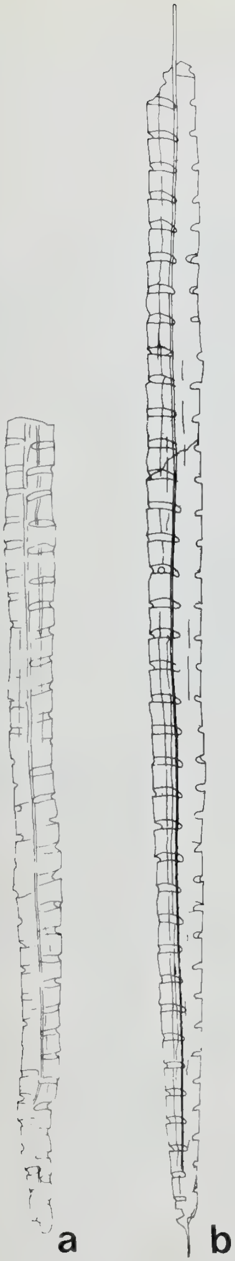

Fig. 6 a, I.G.G.–COAH-SSSP No. 278/5, 1945, a camera lucida drawing of latex cast of the holotype of *Hedrograptus janischewskyi* Obut from the Lower Silurian (Llandovery) of the southern Ural Mountains, U.S.S.R., preserved as a $\frac{3}{4}$-face view impression, × 4; b, I.G.G.–COAH-SSSP No. 278/6, 1945, a 'topotypic specimen' of *H. janischewskyi* 'from the same locality as the holotype and the closest to the type' (Obut, *in litt.* 1984), preserved as a $\frac{3}{4}$-face impression in a light-grey aphanitic limestone with most of the periderm missing, × 4.

Acknowledgements

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Percé, Québec, Canada

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Synopsis

The Ordovician–Silurian boundary in the Percé area occurs within the Matapédia Group. This boundary has not been identified within the Grande Coupe beds, which yield a brachiopod and trilobite fauna with pronounced northwestern European affinities. The Ordovician–Silurian boundary can, however, be recognized within the White Head Formation. The Côte de la Surprise Member is Hirnantian and yields both *Hirnantia* and *Mucronaspis* Communities. The overlying L'Irlande Member is presumed to be totally Silurian, but its basal part has not been positively dated.

Introduction

Southeastern Québec is unique within the North American continent in that it contains two complete sequences near and at the Ordovician–Silurian boundary. A flat-lying sequence of diverse limestones occurs on Anticosti Island (Barnes, this volume), which was originally deposited in a shallow open-marine platform. The Percé sequence is also predominantly limestones, but is decidedly a deeper-platform deposit. This Percé sequence lies within the Appalachian folded belt, at the eastern end of the Aroostook–Percé Anticlinorium, which can be followed from central Maine (USA) to Percé (Ayrton *et al.* 1969), a distance of approximately 500 km. The Aroostook–Percé Anticlinorium in Québec, that is, in Gaspé, lies between the Siluro–Devonian Gaspé–Connecticut River Synclinorium to the north and the Baie des Chaleurs Synclinorium to the south. The Percé area is the most fossiliferous area within the Aroostook–Percé Anticlinorium and, furthermore, the lithostratigraphy there outlined is useful throughout Québec. Thus Percé stands as a local standard for the afore-mentioned anticlinorium.

The Anticosti platform, or the lateral equivalents of it, was probably the source of the carbonates for the Percé sequence. Brachiopods and trilobites are predominantly endemic to each sequence, although corals, conodonts, and ostracodes share some species. Ecological control of these faunas thus appears evident. The Ordovician faunas of the Anticosti and Percé sequences have different faunal affinities: the Anticosti sequence is related to the North American faunas, whereas the Percé faunas have a distinct northwestern European affinity, first recognized by Schuchert & Cooper (1930).

The recognition of the Ordovician–Silurian boundary on Anticosti and around Percé has been treated in detail by Lespérance (1985). A lithostratigraphical and palaeoecological revision of the Early Ashgill to Late Llandovery strata of the Matapédia Group of the Percé area is to be found in Lespérance *et al.* (1987). The lithostratigraphical revision follows the outlines given by Skidmore & Lespérance (1981), while the palaeoecological treatment, relying on the community framework of Boucot (1975), is entirely new. The present contribution will summarize data from Lespérance *et al.* (1981), Lespérance (1985), and Lespérance *et al.* (1987), but will also draw from other sources and unpublished data.

Lithostratigraphical framework

The Aroostook–Percé Anticlinorium in Québec is composed of two main lithostratigraphical sequences: a predominantly carbonate suite termed the Matapédia Group, and a deeper-water, largely turbiditic suite termed the Honorat Group. The Taconic orogeny affected this part of the Appalachians, apparently culminating in the early Caradoc; both the Honorat and Matapédia Groups are younger than early Caradoc. The Honorat does not range into the Silurian (although about a dozen Hirnantian brachiopod localities are known), but the Matapédia

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Group is as young as upper Telychian, on the basis of the conodont *Aulacognathus bullatus* (Nicoll & Rexroad 1969) (as reported by Nowlan 1983), present in the Des Jean Member of the White Head Formation in the Percé area.

Within the immediate vicinity of Percé (Skidmore & Lespérance 1981; Lespérance *et al.* 1987) strata of the Matapédia Group occur in two distinct structural bands. The northeast band is structurally complex, enough so that its total thickness is unknown. It is composed of locally varying proportions of calcilutites and shales, with rare calcarenites, predominantly pelmatozoan-bearing. This northeast band is in fault contact with Cambrian strata to the southwest. The exact lithostratigraphical correlation of these beds with the southwest band (the White Head Formation) is uncertain, which is the main reason why the northeast band of strata has been termed the Grande Coupe beds. Some non-limey shales occur along the sea at Grande Coupe (stream); these have been assigned to the (undivided) Honorat Group, but otherwise, all the Ordovician–Silurian strata of the Percé area are assigned to the Matapédia Group.

The southwest structural band of the Percé area lies with angular unconformity on Cambrian strata. This band is a monoclinal sequence of Ashgill to Llandovery strata which, in turn, are unconformably overlain by the Carboniferous Bonaventure Formation. The lower part of this band is composed of calcareous terrigenous strata and is assigned to the Rouge Member of the Pabos Formation. Above these are limestones, with minor intercalations of fine-grained terrigenous strata, which terminate along the sea at White Head (Cap Blanc); these strata are properly named the White Head Formation. Usage of the term White Head Formation before Skidmore & Lespérance (1981) included the Grande Coupe beds and the Rouge Member of the Pabos Formation, so that care in interpreting previous faunal lists must be exercised.

The stratotypes of the Rouge Member, as well as the four members of the White Head Formation, are all within 6 km of Percé, so that Fig. 1 is representative of the overall stratigraphy. The Rouge Member of the Pabos Formation consists of basal conglomeratic strata and coarse-grained sandstones, followed upward by mud-shales, sandstones, calcarenites, sandy limestones and calcilutites. Terrigenous content decreases upward, and when it reaches less than 50%, this signals the beginning of the White Head Formation.

The basal member of the White Head Formation consists of interbedded thinly bedded calcilutites with thinner interbeds of mudstones, with some calcarenites; these strata form the Burmingham Member. The next member, the Côte de la Surprise, is very predominantly dark green readily-weathering calcareous mudstone. The L'Irlande Member, composed of thin to medium bedded calcilutites and common very thinly bedded mud-shales, as well as rare thin-bedded calcarenites, overlies the Côte de la Surprise Member. Within the middle part of this member are significant clay-shale horizons. The youngest member of the White Head Formation, the Des Jean Member, does not crop out along the type section of the White Head Formation along the sea, and is composed of argillaceous calcilutites, with minor silty and sandy limestones, calcarenites and limestone conglomerates, in fine to very thick beds. The Grande Coupe beds are Ashgill, the Côte de la Surprise Member Hirnantian, and the L'Irlande Member Llandovery. A geological map of the Percé area will be found in Lespérance *et al.* (1987).

Biostratigraphy

Brachiopod-dominated communities, assigned to Benthic Assemblage 4 or 5 (Boucot 1975), dominate the Rouge Member of the Pabos Formation. Extensive brachiopod and trilobite faunas are known from this member (Sheehan & Lespérance 1979), but it is notable that cyclopygid trilobites, as well as the trilobites *Calyptaulax* and *Lonchodomas*, are absent from this member, while *Stenopareia* and *Tretaspis*, on the other hand, are rare; this is in striking contrast with the partly coeval Grande Coupe beds. From a study of encrinurid trilobites, Lespérance & Tripp (1985) suggested that the age of this member was probably Cautleyan.

The Burmingham Member of the White Head Formation is also dominated by brachiopods, which are locally abundant, but their study is difficult because of their preservation in calcilu-

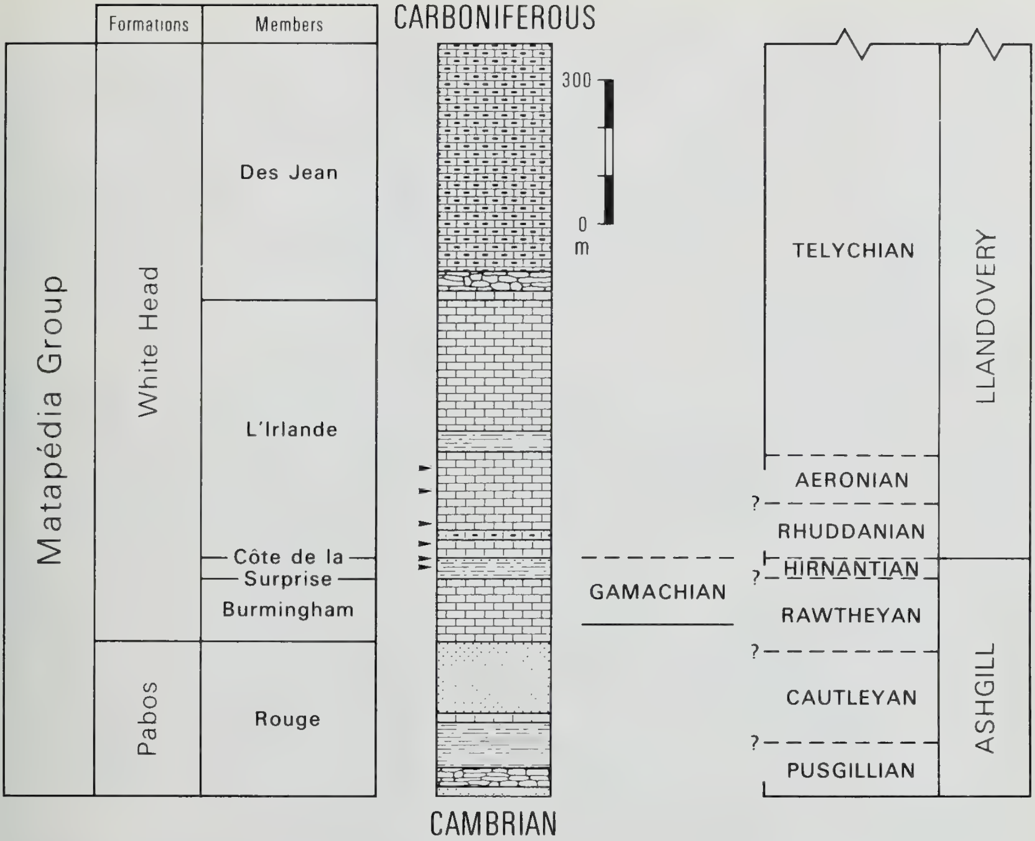
Formations
Members
CARBONIFEROUS
Matapédia Group
White Head
Pabos
Des Jean
L'Irlande
Côte de la
Surprise
Burmingham
Rouge
300
0
m
GAMACHIAN
TELYCHIAN
AERONIAN
RHUDDANIAN
HIRNANTIAN
RAWTHEYAN
CAUTLEYAN
PUSGILLIAN
LLANDOVERY
ASHGILL
CAMBRIAN

Fig. 1 Columnar section of the Pabos and White Head Formations in the Percé area, as taken from the type sections of the various members (covered intervals within type sections filled in by data from adjacent sections). The fossil localities shown within the L'Irlande Member occur along the sea at White Head, where its thickness below the central clay-shale unit is 22 m greater than the one shown for its type section along the Deuxième Rang section. Compiled from data in Lespérance *et al.* (1987). Symbols as in Fig. 2.

tites. Only four trilobite species are known from this member, but corals are present (Bolton 1980). The base of the Gamachian Stage (from Anticosti) is drawn 34 m above the base of this 130 m thick member along the shore at White Head, its stratotype (Lespérance 1985: 841). A Benthic Assemblage 4 has been assigned to this member.

The Des Jean Member fauna is sparsely distributed and dominated by trilobites, notably *Acernaspis* (*Acernaspis*) *primaeva* (Clarke 1908) and *Stenopareia* sp., with infrequent brachiopods. Study of the Des Jean Member and the underlying L'Irlande Member brachiopods is hampered by the preservation in calcilutites and/or calcarenites, and thus most identifications are only precise at the familial level. Nonetheless, these two members have in common *Eospirifer*, a new atrypacid genus and a new athyridacid genus, as well as *Eoplectodonta* cf. *striatacostatus* (Twenhofel 1928); all but the first of these taxa are illustrated in Sheehan & Lespérance (1981: pl. 1). *Oxoplecia* sp. and *Atrypa* sp. are present, but restricted to the Des Jean Member (Lespérance & Sheehan 1981).

Grande Coupe beds

The fauna from the Grande Coupe beds is the best-known fauna from the Percé area, and is the one with the striking northwestern European faunal affinity. No less than 45 different trilobites, 20 brachiopods and 22 cephalopods, to name but these, are known from these beds. The Priest's Road, Grande Coupe and southern flank of Mont Joli (Cooper & Kindle 1936) are its most fossiliferous localities. *Stenopareia perceensis* (Cooper *in* Schuchert & Cooper 1930) [= CSC] and cyclopygid trilobites are abundant, as are locally *Tretaspis clarkei* CSC, *Lonchodomas longirostris* CSC, and the brachiopods *Glyptorthis sublamellosa* CSC, *Sowerbyella gigantea* CSC, *Holtedahlina parva* CSC and *Christiania dubia* CSC. A Benthic Assemblage 6 position is indicated, but with local accumulations of pelagic taxa (cyclopygid trilobites and cephalopods), the *Foliomena* Community (Sheehan & Lespérance 1978), or Benthic Assemblage 4 storm deposits (yielding, notably, colonial corals with encrusted algae).

Hirnantian faunas, or for that matter Silurian faunas, have not been recognized within the Grande Coupe beds.

Côte de la Surprise Member

The stratotype of this member is along the sea at White Head. From a talus slope, approximately in the middle of the member, Lespérance & Sheehan (1976) described the brachiopods and listed other elements present in this fauna: *Dalmanella*? sp., *Eostropheodonta siluriana* (Davidson 1871), *Hirnantia sagittifera* (M'Coy 1851), *Kinnella kielanae* (Temple 1965), *Plectothyrella crassicosta* (Dalman 1828), rare *Phillipsinella parabola s.l.* (Barrande 1846), one pygidium of *Mucronaspis mucronata* (Brongniart 1822), and favositid, cornulitid, conulariid and pelmatozoan taxa. This fauna is a typical *Hirnantia* Community fauna, and assigned a Benthic Assemblage 4 position.

The contact between the Côte de la Surprise Member and the L'Irlande Member is faulted along the sea, and a boundary stratotype has been suggested along the adjacent Deuxième Rang [= Flynn road, Irishtown road] section, where the contact is undisturbed. Here, the uppermost 3 m of the 44 m thick Côte de la Surprise Member is composed of quartz arenites, and these have yielded (Lespérance & Sheehan 1981; Sheehan & Lespérance 1981) abundant brachiopods: an inarticulate, *Dalmanella testudinaria* (Dalman 1828), *Hirnantia sagittifera*, *Kinnella kielanae*, *Eostropheodonta siluriana*, *Plectothyrella crassicosta*, *P.* n. sp., and *Hindella*? sp. (*Hindella*, however, is locally abundant in the Honorat Group west of Percé). This has been assigned a Benthic Assemblage possibly transitional between 3 and 4.

The Côte de la Surprise Member also crops out 17 km west-northwest of Percé (Lespérance 1974; Skidmore & Lespérance 1981) (Fig. 2). The fauna there consists almost entirely of trilobites, with some graptolites, and is a typical Benthic Assemblage 6 fauna. The horizon with the most fossils is between the two covered intervals of Fig. 2; fossils have not been recovered above the uppermost covered interval, nor in the overlying L'Irlande Member. Revision of all previous faunal lists now indicates the presence of: *Brongniartella robusta* (Lespérance 1968), *Cryptolithus portageensis* sp. nov. Lespérance (this volume, p. 370), *Mucronaspis mucronata*, *M. olini* (Temple, 1956), the sponge *Astylospongia praemorsa* (Goldfuss, 1826), a lingulid and a pholidostrophid brachiopod, a bivalve, and the graptolites *Climacograptus normalis s.s.* Lapworth (1877) (J. Riva, personal communication, 1984), and *Orthograptus* sp. This is considered a *Mucronaspis* Community; the presence of graptolites suggests nearness to pelagic (graptolite and other) communities.

L'Irlande Member

Sparsely distributed, often isolated, trilobites and brachiopods occur in the upper three-quarters of the L'Irlande Member, but they are abundant only in infrequent calcarenite beds, often associated with ostracodes. Trilobites are the most abundant taxa in the member, and more specifically *Acernaspis* (*A.*) *primaeva*. The L'Irlande Member has been assigned a Benthic Assemblage 6 position, and named the *Acernaspis* Community (which also includes the overlying Des Jean Member). Although the fauna is sparsely distributed, the total fauna includes species of *Acernaspis* (*Murphycops*), *Bolbineossia*, *Monograptus*, as well as brachiopods (those previously cited as also occurring in the Des Jean Member, as well as *Homoeospira*?, *Streptis* and *Triplesia*), conodonts and trilobites, and is distinctly Llandovery in age. Fossiliferous horizons within and above the clay-shales in the middle of the member are Telychian.

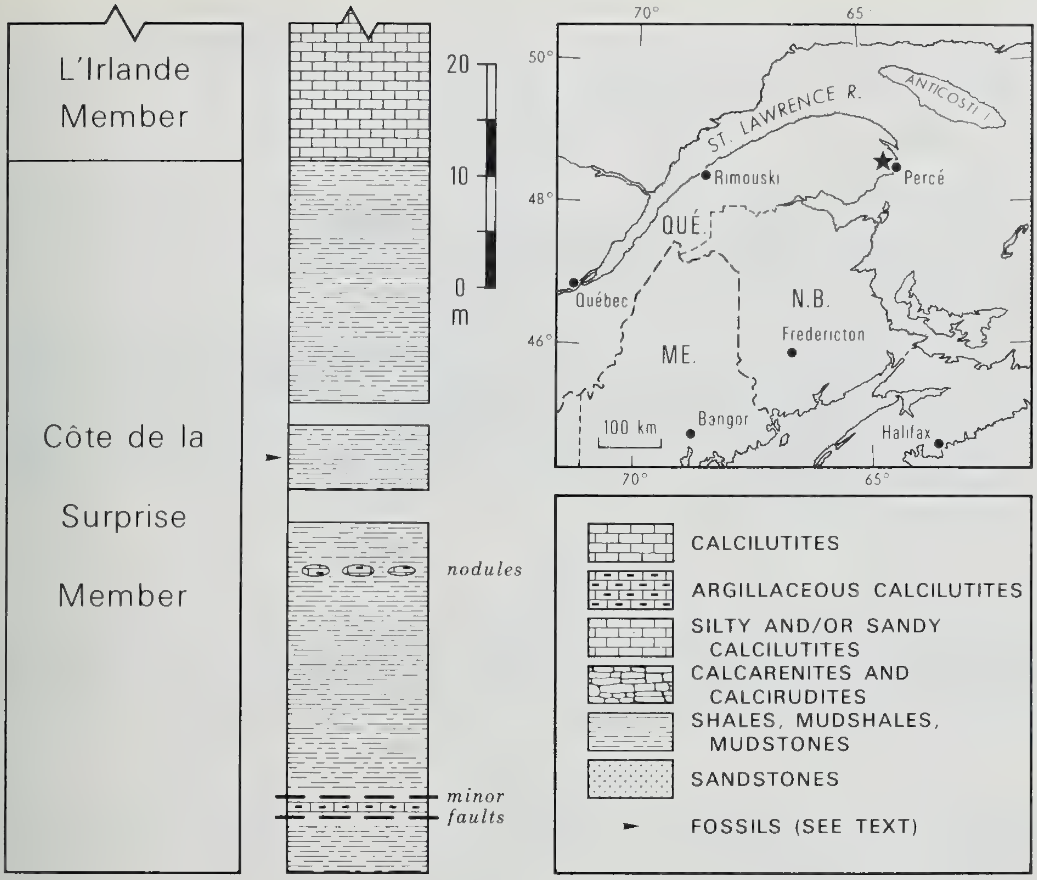
L'Irlande Member
Côte de la Surprise Member
20
10
0
m
nodules
minor faults
70°
65
50°
50
48°
48
46°
46
ST. LAWRENCE R.
ANTICOSTI I.
Rimouski
Percé
QUÉ.
Québec
N.B.
Fredericton
ME.
100 km
Bangor
Halifax
70°
65°
CALCILUTITES
ARGILLACEOUS CALCILUTITES
SILTY AND/OR SANDY CALCILUTITES
CALCARENITES AND CALCIRUDITES
SHALES, MUDSHALES, MUDSTONES
SANDSTONES
FOSSILS (SEE TEXT)

Fig. 2 Columnar section of the Côte de la Surprise Member in the Portage river area (modified from Skidmore & Lespérance 1981). Fossil localities shown by arrowheads are those discussed in the text; numerous others are known. Insert shows location of Percé and the Portage river area (starred); Me.: Maine; N.B.: New Brunswick; Qué.: Québec.

Extensive and closely spaced sampling through the lowest 10 m of the L'Irlande Member along the Deuxième Rang section stratotype has proven fruitless for conodonts (Nowlan 1983: 102).

The L'Irlande Member along the sea at White Head is locally faulted, but, nonetheless, 466 m are present (Lespérance *et al.* 1987). Strata below the middle clay-shale unit (faulted out along the sea) are less fossiliferous than those above, but an extensive trilobite fauna is known 35 m below the clay-shale (62-L31 or locality E of Lespérance *in* Ayrton *et al.* 1969: 479), with *Eoplectodonta* cf. *striatacostatus*, and the new atrypacid and athyridacid genera less than a metre above the trilobites (62-L32). A cephalon of *Acernaspis* sp. occurs 80 m (62-L41; erroneously referred to as a pygidium by Skidmore & Lespérance 1981: 37) below the clay-shales and a pygidium of *Acernaspis*? sp., with *Triplesia* sp., *E.* cf. *striatacostatus* and the two new genera previously quoted (62-L43 of Sheehan & Lespérance 1981: 255) 148 m below the clay-shales. Uncollectable pygidia of *Acernaspis* sp. occur below this last level, some 20–40 m above the base of the member. These are the lowest occurrences of Silurian fossils in the L'Irlande Member in the Percé area.

Lespérance (1985) has attempted to relate the *acuminatus* Zone, the base of the Silurian, to shelly sequences, and has concluded that *Acernaspis* is apparently the only taxon of Silurian aspect, or previously known Silurian distribution, to originate at the *acuminatus* boundary. In view of the presence of the Hirnantian in the topmost Côte de la Surprise Member, the monotonous nature of the L'Irlande Member, and the absence of Ordovician fossils, it appears logical to assign the base of the L'Irlande Member to the Silurian.

Conclusions

Although typical Hirnantian faunas are present in the Percé area, the base of the Silurian cannot be accurately positioned because of the lack of diagnostic graptolites, or, for that matter, other diagnostic taxa. It is surmised that the base of L'Irlande Member is of *acuminatus* Zone age, because *Acernaspis* occurs low in this member.

The Matapédia Group in the immediate Percé area thus consists, in the Ordovician, of deep-water communities (Grande Coupe beds) and shallower communities (Rouge, Burmingham and *Hirnantia* Community of the Côte de la Surprise Member), while the Silurian part reverts to deep-water communities, intermediate between the *Clorinda* and pelagic graptolite communities. The widely accepted glaciation at the end of the Ordovician, although of problematical length (Hambrey 1985), could explain, by rapid eustatic sea-level rise following melting of the ice-caps, the abrupt change from the Côte de Surprise mudstones to the thin-bedded calcilutites of the L'Irlande.

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The Ordovician–Silurian boundary on Manitoulin Island, Ontario, Canada

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Synopsis

The Ordovician–Silurian boundary in southern Ontario is reviewed. Sections on Manitoulin Island have been regarded by earlier workers as representing continuous sedimentation in a shallow carbonate platform environment on the north-east flank of the Michigan Basin. The best section across the boundary, exposed in the Kagawong West Quarry, is described and illustrated. Lithological studies have demonstrated a minor karst development near the systemic boundary. Conodont and macrofossil data demonstrate that the Kagawong Member, Georgian Bay Formation and the lower 15 cm of the overlying Manitoulin Formation are of Richmondian age (Ordovician, Cincinnatian Series). The remainder of the Manitoulin Formation is of Rhuddanian age (Silurian, Llandovery (Anticostian) Series). A hiatus is shown to occur 15 cm above the base of the Manitoulin Formation that represents the Gamachian Stage, Cincinnatian Series and possibly also the latest Richmondian Stage and earliest Rhuddanian Stage. Although the section on Manitoulin Island possesses many of the prerequisites of a boundary stratotype, the hiatus at the systemic boundary ruled it out of consideration as the formal stratotype. It is, however, one of many similar sections in the North American Midcontinent with a hiatus of this proportion at this level which is interpreted as reflecting the eustatic sea level drop in the latest Ordovician related to the north African continental glaciation.

Regional setting

In southern Ontario, undeformed, gently-dipping Ordovician and Silurian carbonates form the eastern margin of the Michigan Basin, affected slightly by the Algonquin Arch (Fig. 1). Over much of this area, the boundary between Ordovician and Silurian strata is a disconformity, but to the north, on Manitoulin Island (Fig. 1), several previous workers have considered it to be conformable with continuous sedimentation. More recent palaeontological and sedimentological work has revealed a paraconformable relationship.

South of the Algonquin Arch (Fig. 1) exposures of the systemic boundary near the base of the Niagara Escarpment reveal a sharp disconformable contact between the Queenston and Whirlpool formations. The Queenston red shales have been generally regarded as continental deposits of the 'Queenston Delta complex' with their widespread distribution being attributed to lowered sea-level caused by the Late Ordovician glaciation (Dennison 1976). A few limestone interbeds low in the Queenston Formation have yielded a marine fauna, including conodonts, brachiopods, and bryozoans with at least the former indicating a littoral community (Barnes *et al.* 1978) and suggesting a Richmondian (Late Ordovician) age. The overlying Whirlpool Formation is a white, cross-bedded sandstone barren of diagnostic fossils, but overlying shales within the Medina Group yield Llandovery fossils. The classic reference section for this area is that of the Niagara Falls gorge.

North of the Algonquin Arch (Fig. 1), the red shales are replaced progressively by shallow water limestone with minor grey shale of the Kagawong Member (30 m) of the Georgian Bay Formation (130 m). On Manitoulin Island the red shales are absent and these Late Ordovician carbonates are overlain by carbonates of the Manitoulin Formation (20 m), regarded as approximately equivalent to the sandstone of the Whirlpool Formation of the Niagara region. These regional stratigraphical relationships are illustrated in Fig. 1.

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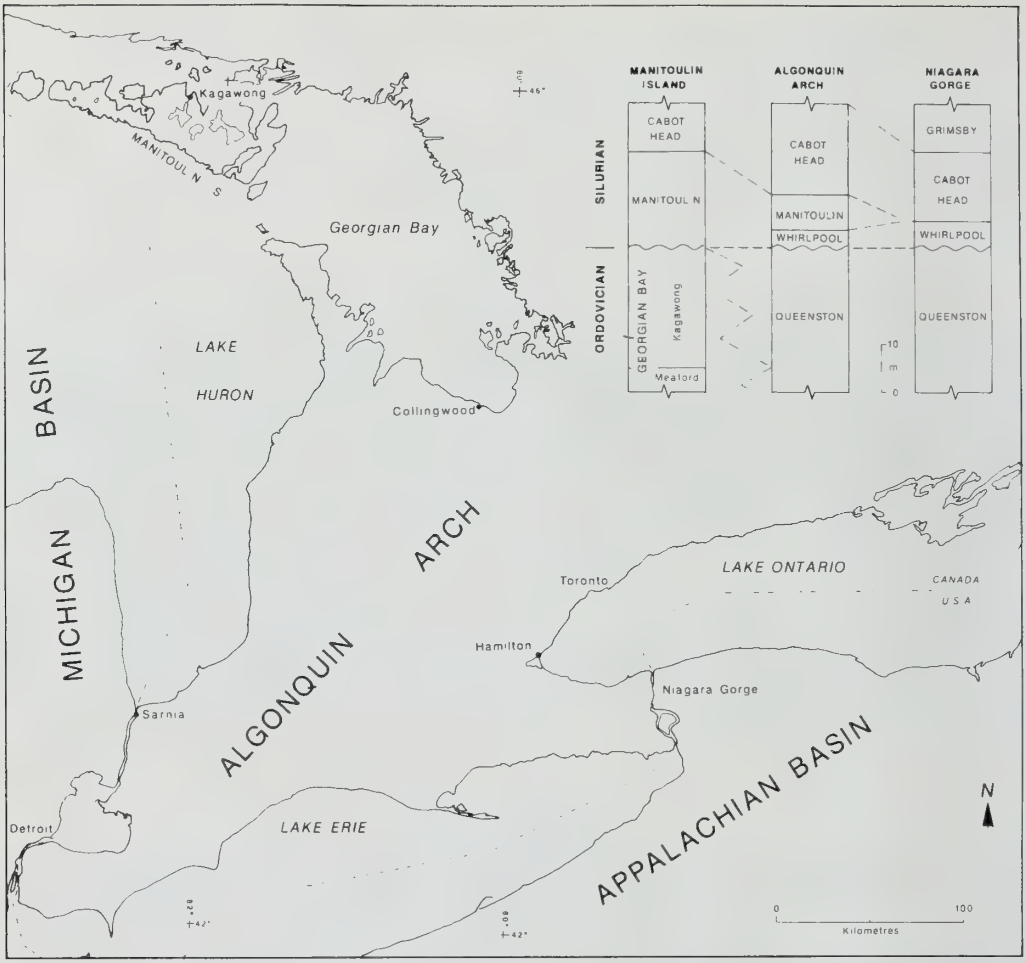

Kagawong
MANITOULIN S
Georgian Bay
LAKE
HURON
Collingwood
MICHIGAN BASIN
ALGONQUIN ARCH
Toronto
LAKE ONTARIO
CANADA
U S A
Hamilton
Niagara Gorge
Sarnia
Detroit
LAKE ERIE
APPALACHIAN BASIN
N
0
100
Kilometres
80°
46°
82°
42°
80°
42°
MANITOULIN ISLAND
ALGONQUIN ARCH
NIAGARA GORGE
SILURIAN
ORDOVICIAN
CABOT HEAD
MANITOULIN
GEORGIAN BAY
Kagawong
Meaford
CABOT HEAD
MANITOULIN
WHIRLPOOL
QUEENSTON
GRIMSBY
CABOT HEAD
WHIRLPOOL
QUEENSTON
10
m
0

Fig. 1 Map of southern Ontario showing Manitoulin Island, main tectonic elements, and generalized stratigraphical successions across the Ordovician–Silurian boundary for Manitoulin Island, Algonquin Arch, and Niagara gorge.

Detailed stratigraphy

On Manitoulin Island, the systemic boundary is best exposed and most accessible at the small, disused Kagawong West Quarry (Figs 2, 3) on Highway 540, 3 km west of Kagawong (Alguire & Liberty 1968, Stop 2; Sanford & Mosher 1978, Stop 10; Telford *et al.* 1981, Stop 14; Kobluk & Brookfield 1982, Stop 6.3). The adjacent roadcut exposes additional strata of the Kagawong Member and the Manitoulin Formation. The following sequence is exposed:

Manitoulin Formation:

6·5 m Dolostone, massive to thick bedded at base, weathering into thin beds separated by irregular shale partings; medium to light brown with grey patches, weathering to a buff colour; medium crystalline; minor vugs in basal 15 cm; abundant fossil debris, especially brachiopods and rugose corals; minor silicification.

0·15 m Dolostone, thin bedded to laminated, argillaceous; medium brown, weathering to a very light brown colour; finely crystalline; beds separated by even shale partings; sharp upper and lower contacts; recessive unit.

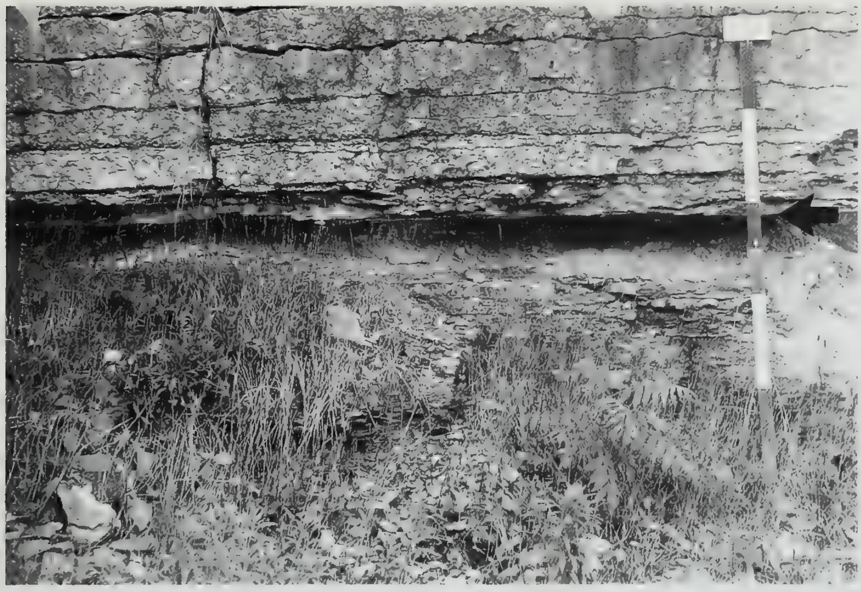


Fig. 2 Kagawong West Quarry showing Kagawong Member, Georgian Bay Formation and Manitoulin Formation. Ordovician–Silurian boundary is drawn (black arrow) at top of 15 cm recessive argillaceous dolostone unit.

Georgian Bay Formation, Kagawong Member:

1·7 m Dolomitic limestone, medium bedded weathering to thin beds; medium grey brown, weathering to blue grey; finely crystalline; poorly fossiliferous, bryozoans and stromatoporoids.

Liberty (1954: 13) and Bolton & Liberty (1954: 28) placed the systemic boundary at the top of the shaly recessive unit, including it within the Kagawong Member. Later Alguire & Liberty (1968: 8) included it in the Manitoulin Formation and considered this sequence to represent continuous sedimentation with no disconformity. Sanford & Mosher (1978: 13) and Sanford *et al.* (1978: 99) from lithological and geochemical evidence placed the systemic boundary 11 cm above the top of the shaly recessive unit, the unconformity probably developing under submarine rather than subaerial conditions. Kobluk (1984) defined two paleokarst surfaces—erosional disconformities below the base and 10 cm above the top of the recessive shaly unit. The lower paleokarst was regarded as at, or very close to, the systemic boundary. Johnson & Telford (1985), however, noted that the disconformable contact between the Manitoulin and Georgian Bay Formations is devoid of scour, rill or other features indicative of extended periods of erosion.

Palaeontology

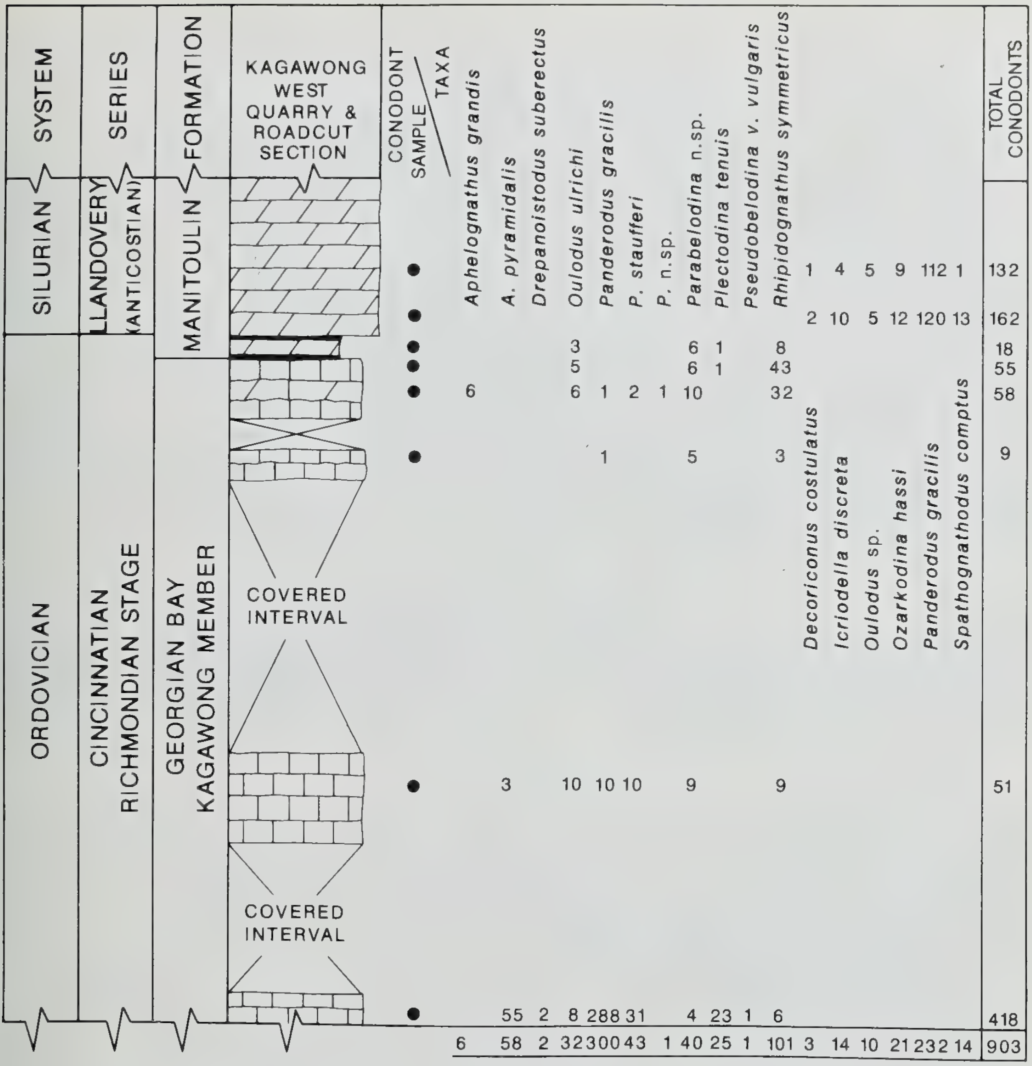
Conodonts. Eight samples from this section, with particular emphasis on the Georgian Bay–Manitoulin formational contact, yielded nearly 1000 conodonts (Fig. 3). This fauna formed part of earlier studies by Tarrant (1977) and Barnes *et al.* (1978). The fauna of the Kagawong Member of the Georgian Bay Formation was listed by Barnes *et al.* (1978: fig. 3) and includes *Aphelognathus grandis* (Kohut & Sweet), *A. pyramidalis* (Branson, Mehl & Branson), *Oulodus ulrichi* (Stone & Furnish), *Panderodus staufferi* (Branson & Mehl), *Pseudobelodina vulgaris* Sweet, *Rhipidognathus symmetricus* Branson, Mehl & Branson. The last species dominates the fauna in the uppermost bed, indicating a littoral environment (e.g. *Rhipiodognathus* community of Barnes & Fåhraeus 1975). The progressive decrease in diversity upwards in the member also suggests upward shallowing. Most taxa are of late Maysvillian to Richmondian age. In the Composite Standard Section for the Middle and Upper Ordovician rocks of the Midcontinent Province, Sweet (1984, Appendix) reports *A. pyramidalis* and *P. staufferi* as restricted to the Richmondian interval. The Kagawong West fauna is herein assigned to the Richmondian *Aphelognathus divergens* Zone. Although several of the taxa range into Gamachian strata on Anticosti Island (McCracken & Barnes 1981: fig. 12), the presence on Manitoulin of *Plectodina tenuis*, *A. grandis* rather than *A.* sp. cf. *A. grandis*, *P. staufferi* rather than *P.* sp. cf. *P. staufferi*, and the absence of *Gamachignathus* spp., suggests a Richmondian rather than a Gamachian age. The fauna may be generally correlative with other Richmondian units such as the Bull Fork and Drakes formations, Cincinnati area (Sweet 1979*a*), the Noix Limestone, Edgewood Group of Missouri (McCracken & Barnes 1982) and the Vauréal Formation of Anticosti Island (Nowlan & Barnes 1981), but biofacies differences between these faunas make precise correlation difficult.

The thin shaly recessive bed, at the base of the Manitoulin Formation, contains a similar fauna with *Rhipidognathus* (Fig. 3). Only *P. gracilis* and possibly *O.* sp. are known to range into Silurian strata elsewhere; no characteristic early Silurian taxa are present. The shaly recessive unit is therefore considered to be of Ordovician (Richmondian) age.

The dolostones of the Manitoulin Formation yielded a conodont fauna (Fig. 3) that includes *Icriodella discreta* Pollock, Rexroad & Nicoll, *Spathognathodus comptus* Pollock, Rexroad & Nicoll s.f., and *Ozarkodina hassi* Pollock, Rexroad & Nicoll. The conodont fauna from the Lower Silurian of southern Ontario, including Manitoulin Island, and northern Michigan was described by Pollock *et al.* (1970), with other documentation by Barnes *et al.* (1978). The lower, but not lowest, part of the Manitoulin Formation thus includes forms indicative of the *Icriodina irregularis* Zone of Pollock *et al.* (1970), who also noted (p. 746) that in some sections 'the oldest parts of the Manitoulin ... seems to correspond with the pre-*Icriodina irregularis* Zone in the Midwest ... and with the lower part of Walliser's (1964) Bereich I.' *I. discreta* and *O. hassi* are known from earliest Silurian strata, Menierian Stage of Barnes (in press), in the Anticosti Island sections that are continuous across the Ordovician–Silurian boundary although *S. comptus* is absent (McCracken & Barnes 1981: fig. 12; Barnes, this volume). Herein, the Manitoulin Formation is assigned to the *Icriodella discreta–I. deflecta* Zone of Aldridge (1972). In the Manitoulin section, there is therefore no evidence of the latest Ordovician conodont Fauna 13 characteristic of the Gamachian Stage as described by McCracken & Barnes (1981) from Anticosti Island. Other sections in the Midcontinent in North America also lack this interval, e.g. the Cincinnati area (Sweet 1979*a*; Sweet *et al.* 1971; Sweet 1984), the Noix Limestone and Bowling Green Dolomite of the Edgewood Group, Missouri (McCracken & Barnes 1982), and elsewhere in the western Midcontinent (Sweet 1979*b*), and the Hudson Bay region (LeFèvre *et al.* 1975). McCracken & Barnes (1981) attributed this pattern to the latest Ordovician (Gamachian) regression, induced by the north African glaciation, which restricted areas of continuous sedimentation to subsiding marginal cratonic basins or non-eroding oceanic basins.

Macrofossils. The general fauna of the Kagawong Member, Georgian Bay Formation, as detailed by Caley (1936), suggests the inclusion of these carbonates within the standard North American Richmondian Stage. Within the upper 5 m, only *Stromatocerium*, *Tetradium* and poorly preserved undiagnostic bryozoans, bivalves and gastropods have been identified. According to Copper (1982: 680), 'the post-Richmondian Ellis Bay *Spirigerina–Hindella* faunas of Anticosti Island are absent, suggesting an interval of erosion or non-deposition'.

The fauna of the overlying Lower Silurian Manitoulin Formation is scattered throughout with concentrations confined to the uppermost beds (Bolton 1966, 1968). Characteristic forms include the corals *Paleofavosites asper* (d'Orbigny), *Palaeophyllum williamsi* Chadwick, cystoid *Brockocystis tecumseth* (Billings), brachiopods *Resserella eugeniensis* (Williams), *Mendacella* sp., '*Orthorhynchula*' *bidwellensis* Bolton, *Zygospiraella planoconvexa* (Hall), *Sypharatrypa* (?) *laticorrugata* (Foerste), *Eospirigerina parksi* (Williams), and *Dolerorthis* sp. An early Llandovery (Anticostian) pre-C_3 age, within the '*Coelospira*' *planoconvexa*–'*Atrypa*' *laticorrugata* Zone of



Conodont sample	*Aphelognathus grandis*	*A. pyramidalis*	*Drepanoistodus suberectus*	*Oulodus ulrichi*	*Panderodus gracilis*	*P. staufferi*	*P.* n.sp.	*Parabelodina* n.sp.	*Plectodina tenuis*	*Pseudobelodina v. vulgaris*	*Rhipidognathus symmetricus*	*Decoriconus costulatus*	*Icriodella discreta*	*Oulodus* sp.	*Ozarkodina hassi*	*Panderodus gracilis*	*Spathognathodus comptus*	Total conodonts
●												1	4	5	9	112	1	132
●												2	10	5	12	120	13	162
●				3				6	1		8							18
●				5				6	1		43							55
●	6			6	1	2	1	10			32							58
●					1			5			3							9
●		3		10	10	10		9			9							51
●		55	2	8	288	31		4	23	1	6							418
	6	58	2	32	300	43	1	40	25	1	101	3	14	10	21	232	14	903

Fig. 3 Section at Kagawong West Quarry showing distribution of conodont species in upper Kagawong Member, Georgian Bay Formation and in lower Manitoulin Formation, across the Ordovician–Silurian boundary.

Ehlers & Kesling (1962: 7), is assigned to the Manitoulin Formation. In the Kagawong West Quarry, proper *Brockocystis tecumseth* was discovered near the base of the Manitoulin Formation and the first brachiopod concentration was located 1 m above the base. Copper (1978: 51) reported 'the atrypoid *Zygospiraella*, an index genus from earliest Llandoverian (A) strata on the Siberian platform and in the Baltic area is common' from the basal few centimetres of the Manitoulin Formation (above the recessive shaly dolostone bed). A Llandovery A age is also assigned to the Manitoulin and overlying Cabot Head formations by Johnson (1981).

Summary

In the classic Niagara gorge section of southern Ontario there is an undisputed disconformity between late Ordovician and early Silurian strata. To the north, on Manitoulin Island, several previous workers have argued for continuous sedimentation within a carbonate sequence across the systemic boundary. Recent studies of the last decade on both conodonts and macrofossils now indicate a paraconformable relationship with the systemic boundary lying 15 cm above the base of the Manitoulin Formation and associated with subtle paleokarst development. The Kagawong Member of the upper Georgian Bay Formation and the basal 15 cm of the Manitoulin Formation are assigned to the *Aphelognathus divergens* Zone of the Richmondian Stage, Cincinnatian Series. The Manitoulin Formation is assigned to the *Icriodella discreta–Icriodella deflecta* Zone and the Llandovery A, i.e. Rhuddanian Stage (Menierian Stage), Llandovery (Anticostian) Series. The hiatus within the lower Manitoulin Formation therefore represents the Late Ordovician Gamachian Stage and possibly the latest Richmondian and earliest Rhuddanian (Menierian) as well. This hiatus is regionally extensive across the Midcontinent (Barnes *et al.* 1981; Ross *et al.* 1982) and is interpreted as a result of eustatic sea-level drop related to the Late Ordovician continental glaciation in north Africa.

The Kagawong West Quarry section is well exposed, undeformed with low burial temperatures of CAI 1·5 (Legall *et al.* 1982) and with strata dipping at less than five degrees, moderately fossiliferous, readily accessible, and has other qualities expected of a boundary stratotype. However, even as the best potential section in southern Ontario, the recent demonstration through detailed faunal and lithologic studies of a hiatus at the systemic boundary ruled out this section as the boundary stratotype.

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Preliminary report on Ordovician–Silurian boundary rocks in the Interlake area, Manitoba, Canada

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Synopsis

Both Ashgill and early Llandovery rocks are represented in both surface outcrop (Stonewall Quarry) and the subsurface of Manitoba, but there is no definite evidence of continuous sedimentation through the boundary period.

The Interlake area of central Manitoba and its northwestward extension to eastern Saskatchewan (Fig. 1) provides the only outcrop area for the Lower Palaeozoic strata of the Williston Basin, and the only Lower Palaeozoic outcrops between Hudson Bay and the western Cordillera. Unfortunately, outcrops are sparse and expose only limited stratigraphical intervals, so that it is not possible at present to propose a definitive locality for the Ordovician–Silurian boundary there. No single outcrop area is at present known which exposes completely the required stratigraphical interval. Nevertheless, because of the critical location of the Manitoba outcrop belt, the following will present a brief summary of data relevant to the delineation of the boundary.

Stearn (1953, 1956), on the basis of detailed faunal studies, placed the Stonewall Formation in the Ordovician and placed the Ordovician–Silurian boundary at the contact between the Stonewall Formation and the overlying Fisher Branch dolomite of the Silurian Interlake Group. However, because of erosion of the uppermost beds, the type section of the Stonewall Formation at the Stonewall Quarry is incomplete. At the time of Stearn's studies, firm correlation with the complete subsurface sequence had not been established. Subsequently, Porter & Fuller (1959) established a subsurface reference section for the Stonewall Formation, based on correlation of regional marker horizons (B. A. Morriseau, 8-20-90-6W; 875′-920′). Detailed correlations between the Morriseau well and the Stonewall Quarry (about 72 km to the east) indicate that, at the Stonewall Quarry, the uppermost 4 to 6 m of the Stonewall beds, including a prominent medial arenaceous-argillaceous marker bed (t-horizon) has been eroded. Brindle (1960), from subsurface faunal studies, suggested that the Ordovician–Silurian boundary falls within the Stonewall Formation, rather than at the top, and may be marked by the medial arenaceous bed. It must be noted that marker beds at the top, middle and bottom of the Stonewall Formation can be correlated through almost the entire Williston Basin, indicating little or no stratigraphical discontinuity at the Ordovician–Silurian boundary.

Preliminary results of conodont studies (C. R. Barnes, personal communication) indicate an Ordovician–Richmondian (Ashgill) age for the Stonewall Quarry beds. Also, a possible late Lower Llandovery fauna was obtained from a core hole drilled near the outcrop belt north of Grand Rapids. Exact correlation of this core hole with the surface section is uncertain, but it appears that the sampled interval may be upper Stonewall, and the upper Stonewall beds may, at least in part, fill the apparent gap between the lower Stonewall beds of Ashgill age and the Middle Llandovery Fisher Branch beds.

Recent stratigraphical core hole drilling in the Interlake outcrop belt, and mineral exploration drilling in the area north and west of Grand Rapids, have obtained a number of cores for the Fisher Branch–Stonewall–Stony Mountain succession, so that the complete lithological sequence through the Ordovician–Silurian boundary interval is now available. Also, recent geological mapping has outlined several new outcrops that may expose this interval. Although no systemic boundary outcrop can be defined with certainty, two newly accessible occurrences may possibly include the boundary zone, but precise faunal data for these outcrops are not yet

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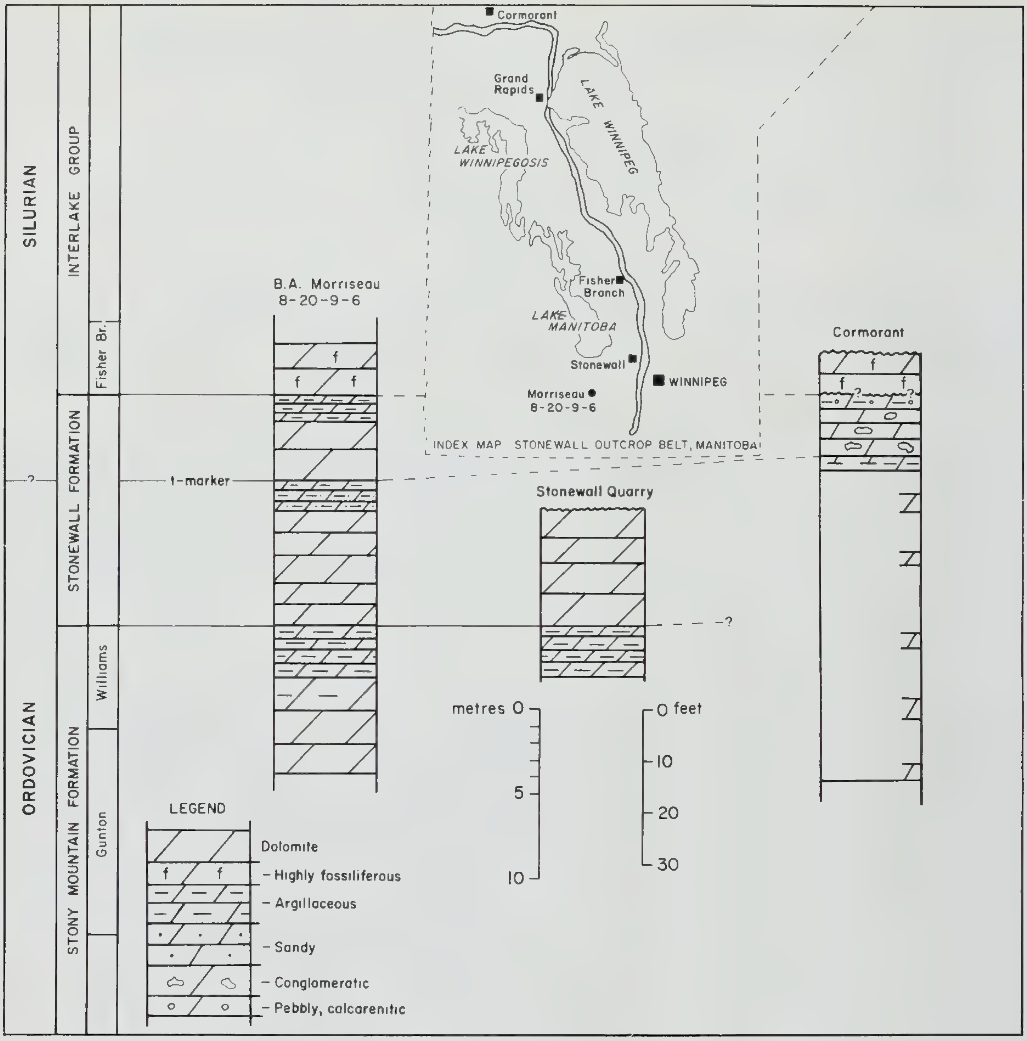
SILURIAN
ORDOVICIAN
INTERLAKE GROUP
STONEWALL FORMATION
STONY MOUNTAIN FORMATION
Fisher Br.
Williams
Gunton
?
t-marker
B.A. Morriseau
8-20-9-6
Cormorant
Grand Rapids
LAKE WINNIPEG
LAKE WINNIPEGOSIS
Fisher Branch
LAKE MANITOBA
Stonewall
WINNIPEG
Morriseau
8-20-9-6
INDEX MAP STONEWALL OUTCROP BELT, MANITOBA
Stonewall Quarry
Cormorant
?
metres 0
5
10
0 feet
10
20
30
LEGEND
Dolomite
- Highly fossiliferous
- Argillaceous
- Sandy
- Conglomeratic
- Pebbly, calcarenitic

Fig. 1 Correlation of the Stonewall Formation and adjacent rocks in the Interlake area, Manitoba, Canada (in part after Porter & Fuller 1959). Correlation with the subsurface is also shown.

available. A thin sequence of dolomites, including an argillaceous marker bed believed to be the mid-Stonewall (t-horizon) marker, is exposed at the parking lot for the Manitoba Hydro powerhouse at Grand Rapids, but the remaining stratigraphical exposure is minimal.

A large bedrock hill south of the village of Cormorant (approx. Sec. 14, Tp. 60, Rge. 22 WPM), on the south shore of Cormorant Lake, is traversed by a recent extension of Provincial Road 287. This hill is believed to comprise an outlier of the Stonewall Formation, although exposure is by no means complete (Fig. 1). Good exposures occur in a roadcut at the top of the hill, in a small quarry near the top, and in a number of scattered natural outcrops on the slopes of the hill. Total topographic relief (partially exposed stratigraphical section) is 33 m, and the estimated Stonewall thickness is only about 10·6 m. A preliminary examination shows, at the top of the hill, a 2–3 m cap of massive to nodular bedded, buff mottled, variably fossiliferous dolomite with numerous corals and minor brachiopods and gastropods, but no recognizable

Virgiana decussata (the diagnostic fossil of the Fisher Branch Formation). These beds have not yet been identified palaeontologically, but on the basis of lithology are believed to be Fisher Branch Formation (Middle Llandovery). These beds overlie sharply, and with apparent slight unconformity, a pebbly argillaceous marker bed (0·9 m), which in turn is underlain by fine-grained dense conglomeratic dolomite (2·87 m). This in turn overlies a 0·64 m reddish grey dolomitic shale and argillaceous dolomite (possible t-marker?) which passes downward to microcrystalline dense dolomites. The conglomeratic beds are believed to be stratigraphically equivalent to similar dolomites described by Stearn for an outcrop on P.T.H. 10 near Rocky Lake, 26·7 miles (42·6 km) north of The Pas (Stearn 1956: 13). Stearn reported an Ordovician fauna from these strata, suggesting that, at this locality and at Cormorant, a portion of the Upper Stonewall may be missing because of non-deposition or pre-Fisher Branch (Middle Llandovery) erosion.

Core-hole drilling and microfossil studies for the Cormorant section and for the Stonewall area, planned for 1986–87, may permit more precise determination of the Ordovician–Silurian boundary in Manitoba. It should be noted that the conglomeratic beds occurring in the Stonewall Formation in central Manitoba (e.g. the Cormorant area) are not known in southern Manitoba, where the Stonewall beds are slightly thicker and possibly comprise a more complete, but not completely exposed, Ordovician–Silurian boundary sequence.

The summary faunal list for the Stonewall Formation is as follows:

Upper Stonewall fauna (after Brindle 1960 for Saskatchewan subsurface):
  Above t-marker: streptelasmid, *Favosites* cf. *favosus* Goldfuss, *Syringopora* sp., bryozoan.
  Below t-marker: *Halysites* (*Catenipora*) *gracilis* Hall, ?*Oepikina stonewallensis* Stearn. *Spathognathus manitoulinensis* (Pollock, Rexroad & Nicoll)—C. R. Barnes (pers. comm. 1975).

Lower Stonewall fauna (Stonewall Quarry section—after Stearn 1956): *Kochoceras* cf. *productum*, *Antiplectoceras shammattawaense*, *Paleofavosites capax*, *P. okulitchi*, *Tryplasma gracilis*, *Angopora manitobensis*, *Beatricea regularis*, *Megamyonia nitens*, ?*Oepikina stonewallensis*, *Ephippiorthoceras minutum*, *Metaspyroceras meridionale*, *Bickmorites insignis*.
  (after C. R. Barnes 1975, pers. comm.): *Belodina profunda* (Branson & Mehl), *Rhipidognathus symmetrica discreta* Bergström & Sweet, *Panderodus staufferi* (Branson, Mehl & Branson).

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The Ordovician–Silurian boundary in the Rocky Mountains, Arctic Islands and Hudson Platform, Canada

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Synopsis

The Ordovician-Silurian Boundary is developed within sequences of platform carbonates at Pedley Pass (southeastern British Columbia) and in the Kaskattama well (northeastern Manitoba). At Snowblind Creek (Arctic Islands), the boundary is documented within a transitional facies between platform carbonates and basinal rocks, but access to the locality is difficult and expensive. Further detailed palaeontological studies are needed to establish the precise position of the boundary at all three localities.

The Rocky Mountains

Silurian carbonates are widespread in parts of the Rocky Mountains (1400 km long, 50–140 km wide) and a graptolitic facies is locally present in the northwestern and west-central parts. Access is expensive except close to the very few roads. In the graptolitic facies (Road River Group), the Ordovician–Silurian boundary interval has not been studied in detail. Exposures are not good and unconformities are present within or below the Llandovery part of the sequence. The Ordovician *ornatus* Zone and the Silurian *cyphus* Zone are well documented (Cecile & Norford 1979; Jackson *et al.* 1965; Davies 1966) and taxa identified by Davies may indicate some of the intervening *persculptus*, *acuminatus*, *atavus* and *acinaces* Zones.

The carbonate facies consists of resistant dolomites almost throughout the Rocky Mountains. North of Peace River an unconformity is present below Silurian dolomites of the Nonda Formation. South of the Peace, the Beaverfoot Formation appears to span latest Ordovician and most of Llandovery time.

The section at Pedley Pass is typical of many in southeastern British Columbia, except that access is simple and inexpensive. The locality is within a carbonate platform, a considerable distance inboard of the platform-front. Exposure is excellent along a steep ridge above the timberline, and complete through more than 500 m of Upper Ordovician and Lower Silurian limestones and dolomites. Retreat of glaciers was relatively recent and the rocks are essentially unweathered. The terrane is folded and thrust but structure is simple within the thrust plates, with moderate dip parallel to the ridge. Disconformities have not been recognized within the boundary interval, but discontinuities could be present within the sequences of shallow water carbonates. Conodont alteration indices (CAI) of 4 are known from just above the Beaverfoot Formation near Pedley Pass (Goodarzi & Norford 1985: 1091, sample D) and thus the rocks of the boundary interval have high thermal maturity and are quite unsuitable for palaeomagnetic and many geochemical studies.

At Pedley Pass, 130 m of poorly fossiliferous dolomites separates an Upper Ordovician coral and brachiopod fauna (*Bighornia–Thaerodonta* Fauna of Ashgill age) from the lowest brachiopods (*Nondia* sp.) and corals (*Rhegmaphyllum* sp., *Streptelasma* sp.) confidently dated as Silurian (*Eostropheodonta* Zone, part of *Virgiana* fauna, upper Lower to Middle Llandovery). Macrofossils are present in the intervening rocks but are poorly preserved. Conodont studies of these beds have not been completed, but preliminary data (T. T. Uyeno *in* Norford 1969: 39) from a corresponding interval at Mount Sinclair, 25 km north of Pedley Pass, indicate that the Ordovician–Silurian Boundary lies somewhere within the upper 75 m of the poorly fossiliferous interval at Pedley Pass.

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Thus, the Beaverfoot Formation seems to show sedimentation across the Ordovician–Silurian Boundary but the problems are those of precisely locating the boundary and the high thermal maturity (CAI 4) of the rocks. The region is not suitable for a stratotype of more than local application.

The Arctic Islands

The Arctic Platform and the Inuitian Orogen comprise a vast region (2000 by 1000 km) in which Ordovician and Silurian rocks are widely distributed, both in outcrop and subsurface. Exposures are mostly good, but logistic dependency on aircraft makes access expensive and then only possible during the short summer. A carbonate shelf is bounded to the northwest by a graptolitic facies, locally stratigraphical sections show the interfingering of the two facies in great detail, for example, along Snowblind Creek, Cornwallis Island (Thorsteinsson 1959). Broad open folds characterize the structure in most of the Arctic Platform; thermal maturities are low on Cornwallis Island (Conodont Alteration Indices 1 to 2, Uyeno 1981 and *in* Goodarzi & Norford 1985: 1091, sample B). Macrofossils are not common in the carbonate facies, but the graptolitic facies is very fossiliferous, locally with exquisite preservation of graptolites in full relief within limestone nodules. Palaeontological studies of both macrofossils and microfossils are only at a reconnaissance level at present, but the region has great promise for the achievement of detailed correlations of zonal schemes based on various phyla.

Carbonates of the Allen Bay Formation, the Baillarge Formation and correlative rocks contain corals, cephalopods, brachiopods, gastropods, trilobites and receptaculitids. Ashgill faunas resemble those of northwestern Greenland and the Hudson Platform. Conodont faunas indicate Fauna 12 of the United States with the same fauna present in latest Caradoc rocks; Fauna 13 may also be present below conodont faunas indicative of the mid-continent Lower Silurian *kentuckyensis* Zone (Ryley 1984). Very early Silurian macrofaunas have not yet been collected from these formations, and, similarly, the conodont faunas are poorly known.

Most probably, all of latest Ordovician and earliest Silurian time is represented within the Cape Phillips and Ibbett Bay Formations of the graptolitic facies. However, the graptolite faunas have not yet been described taxonomically and the presence of the *pacificus*, *extraordinarius*, *persculptus* and *acuminatus* Zones have not been established. Cephalopods, radiolarians, sponge fragments, ostracodes, polychaetes and trilobites are associated with the latest Ordovician graptolite faunas and allow correlation into the carbonate facies.

Thus, the Late Ordovician and Early Silurian macrofaunas and microfaunas have yet to be described, but the intricate facies relations of carbonates and graptolitic rocks make it a region of international importance for the discrimination of the Ordovician–Silurian Boundary. The section at Snowblind Creek on Cornwallis Island is eminently suitable as a key section for intercontinental correlations except for its difficult access. The variety of fossil groups within the graptolite zones provides for detailed correlation of shelly benthic zones with the standard graptolite zonation.

The Hudson Platform

The Hudson Platform is a large remnant (1600 by 1000 km) of a sequence of Palaeozoic carbonates and evaporitic rocks that once covered much of the Canadian Shield. The platform now floors Hudson Bay, but the rocks extend onshore in the Hudson Bay and James Bay Lowlands to the south and on Southampton, Coats and Mansel Islands to the north. Access to all of these areas is difficult and costly. Outcrop is very sparse in the Lowlands and limited to the major rivers and some intertidal regions; exposures are less rare in the northern islands but stratigraphical sections are few and incomplete. The rocks are essentially flat-lying with rare faults. Thermal maturities are low. A number of wells have been drilled in the Lowlands and offshore in the central regions of Hudson Bay; these provide the best stratigraphical sections and several (including Sogepet–Aquitaine Kaskattama Province No. 1) took continuous slim core through the Ordovician–Silurian Boundary.

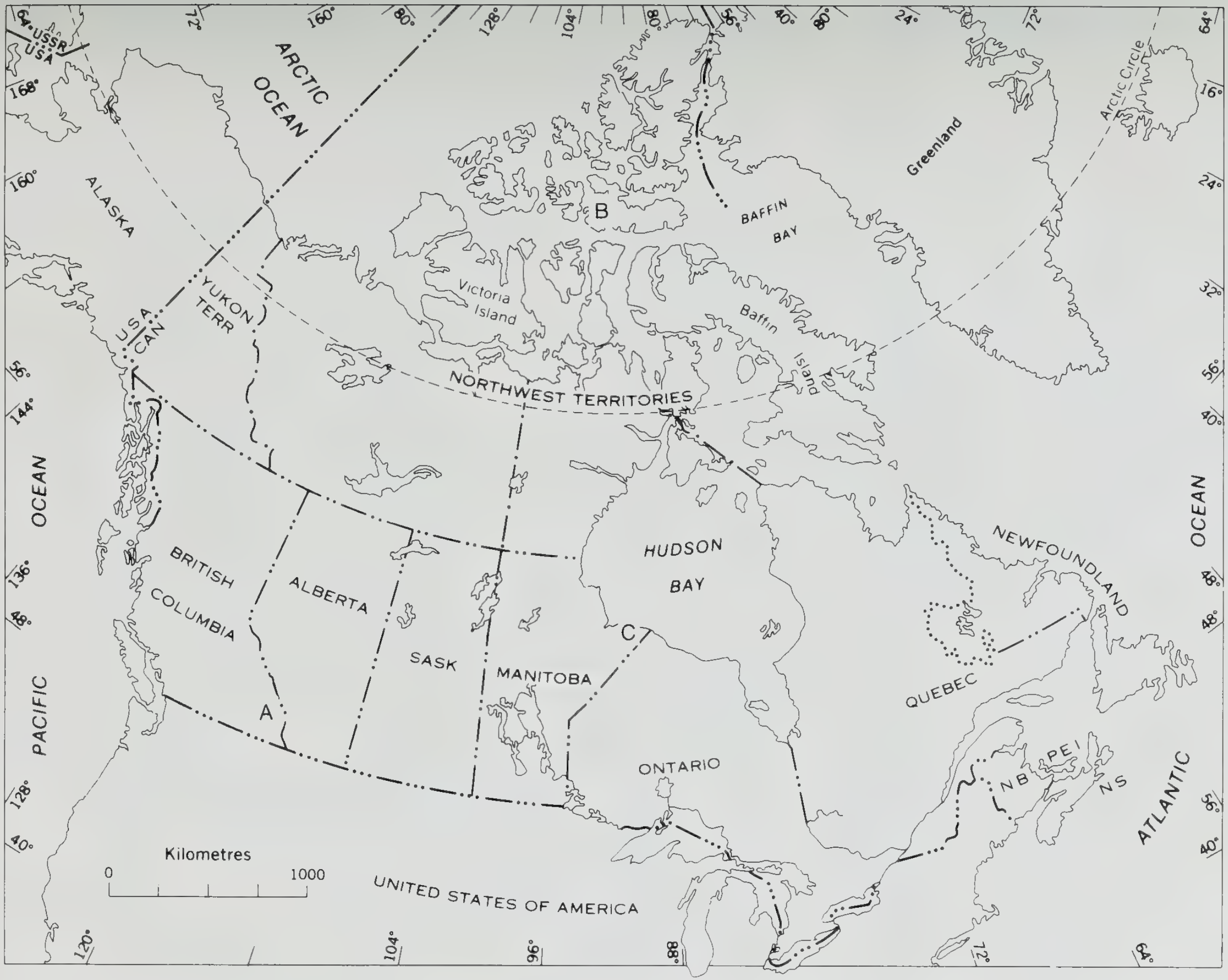
64°
USSR
USA
168°
72°
160°
80°
128°
104°
80°
56°
40°
80°
24°
72°
64°
16°
24°
32°
56°
40°
48°
48°
56°
40°
ARCTIC OCEAN
ALASKA
USA
CAN
YUKON TERR
Victoria Island
B
BAFFIN BAY
Greenland
Arctic Circle
Baffin Island
NORTHWEST TERRITORIES
56°
144°
OCEAN
BRITISH COLUMBIA
ALBERTA
SASK
MANITOBA
HUDSON BAY
C
A
136°
48°
PACIFIC
128°
40°
ONTARIO
QUEBEC
NEWFOUNDLAND
OCEAN
NB
PEI
NS
ATLANTIC
Kilometres
0
1000
UNITED STATES OF AMERICA
120°
104°
96°
88°
72°
64°

Fig. 1 Locality Map. A, Pedley Pass Section, Rocky Mountains of southeastern British Columbia. B, Snowblind Creek Section, Cornwallis Island, Northwest Territories. C, Sogepet–Aquitaine Kaskattama Province No. 1 Well, Hudson Platform, northeastern Manitoba.

Two sets of nomenclature have been used for an interval of shallow water dolomites between the Churchill River Group (Ashgill) and an unconformity at the base of the Severn River Formation (basal beds high Lower or Middle Llandovery). The Port Nelson Formation (Savage & Van Tuyl 1919; Norford 1971) is based on an outcrop on Nelson River, northern Manitoba; the Red Head Rapids Formation (Nelson 1963, 1964; Sanford 1974; Heywood & Sanford 1976) is based on two outcrops on Churchill River in the same region. The relations are uncertain between these three outcrops and an interval recognized in the subsurface between the Churchill River Group and the Severn River Formation. Nomenclatorial priority suggests the use of the term Port Nelson Formation. The interval reaches 35 m in the subsurface of northern Manitoba and is more than 60 m thick on Southampton Island, where some additional younger strata may be present beneath the sub-Severn River unconformity.

The Silurian Severn River Formation rests on 32 m of the Port Nelson Formation in the Kaskattama well (Norford 1970). The contact is apparently conformable in the well but elsewhere there is evidence of an erosion surface beneath the Severn River Formation. The lower part of the Severn River Formation can be dated as late Early or Middle Llandovery, primarily on the presence of *Virgiana decussata* (Whiteaves). In the well, the Port Nelson Formation consists of dolomites and dolomitic limestones with mudstone partings and nodules, isolated crystals and thin beds of anhydrite and locally halite. In the Kaskattama well, corals and brachiopods in the basal 11 m indicate a very late Ashgill age and correlation with the lower

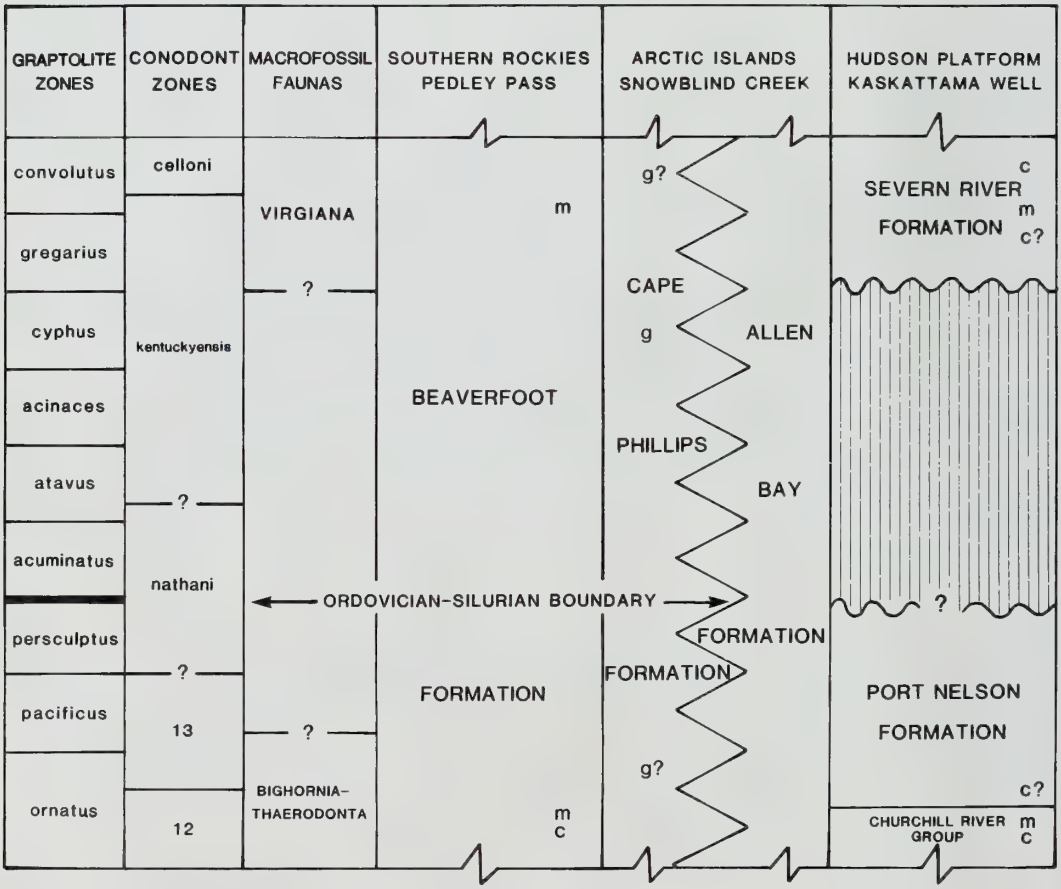
GRAPTOLITE ZONES
CONODONT ZONES
MACROFOSSIL FAUNAS
SOUTHERN ROCKIES PEDLEY PASS
ARCTIC ISLANDS SNOWBLIND CREEK
HUDSON PLATFORM KASKATTAMA WELL
convolutus
gregarius
cyphus
acinaces
atavus
acuminatus
persculptus
pacificus
ornatus
celloni
kentuckyensis
?
nathani
?
13
12
VIRGIANA
?
?
BIGHORNIA–THAERODONTA
m
BEAVERFOOT
FORMATION
m
c
g?
CAPE
g
PHILLIPS
FORMATION
g?
ALLEN
BAY
FORMATION
ORDOVICIAN–SILURIAN BOUNDARY
c
SEVERN RIVER FORMATION
m
c?
?
PORT NELSON FORMATION
c?
CHURCHILL RIVER GROUP
m
c

Fig. 2 Correlation diagram of the three areas.

and middle part of the Stonewall Formation of southern Manitoba. The upper 21 m contains only fragmentary fossils in the well, but sparse conodonts (Conodont Alteration Index 1 to 1·5) have been recovered from outcrop 5·5 m below the top of the Port Nelson Formation in its type section. T. T. Uyeno has identified *Panderodus* cf. *P. simplex* Branson & Mehl s.f. and tentatively dates the horizon as early Llandovery, but comments that the form shows some transitional features to those of the Middle and Upper Ordovician form-species *Panderodus compressus* Branson & Mehl.

Thus, in the Hudson Platform the Ordovician–Silurian Boundary lies either within the Port Nelson Formation or within a regional unconformity below the Severn River Formation. The sediments that formed the Port Nelson Formation were inhospitable to animal life, and although one can hope for more refined dating of the upper beds and thus more precise positioning of the Boundary, the region is not suitable for a stratotype of more than local application.

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Ordovician–Silurian boundary, northern Yukon, Canada

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Synopsis

The Ordovician–Silurian boundary is described from three graptolite and conodont-bearing sections of northern Yukon. Upper Ordovician graptolite biostratigraphical units comprise the *Dicellograptus ornatus*, *Pacificograptus pacificus* and tentatively the *Glyptograptus persculptus* zones; that of the conodonts being the *Amorphognathus ordovicicus* Biozone and the North American Fauna 12. A stratigraphical hiatus between the *P. pacificus* Zone and the *G. persculptus* Zone?, and probably equivalent to the *Diplograptus bohemicus* and *Climacograptus*? *extraordinarius* graptolite Zones, and to North American conodont Fauna 13, appears to be present everywhere in the region.

Introduction

The presence of excellently exposed graptolite-bearing sequences in the Richardson and Ogilvie mountains of northern Yukon has been recognized for more than 20 years. Graptolitic strata of the Road River Formation are known to be widely distributed throughout the northern Cordillera of Canada and adjacent Alaska (e.g. Lenz & Perry 1972; Lenz 1972, 1982; Churkin & Brabb 1965).

For the purpose of this paper, three key sections are discussed; these are Peel River, Pat Lake and Blackstone River, the first in the Richardson Mountains, the latter two in the Ogilvie Mountains (Figs 1, 2). The Peel River section is chosen because the Ordovician–Silurian boundary beds are completely exposed and are well studied, and are defined on both graptolites and conodonts; Pat Lake contains a thick conodont and shelly fauna-bearing limestone of probable latest Ordovician age in an otherwise entirely graptolitic sequence; and Blackstone River is a much thicker boundary sequence containing both graptolites and conodonts. These sequences have already been discussed in Lenz & McCracken (1982).

The three sections discussed are in remote and isolated areas of northern Yukon, the field season is relatively short, seldom more than two and a half months, and the cost of access is high. Access to any of the three localites is via regular scheduled aircraft service to Whitehorse in southern Yukon, and then to the villages of either Mayo or Dawson City in central Yukon (Fig. 1), and by privately chartered helicopter thereafter. Weather in the region can vary considerably, but is generally pleasant in July and early August.

Stratigraphy

The Road River Formation, the type of which is in the Richardson Mountains (Jackson & Lenz 1962), is a thick basinal sequence of dominantly dark grey to black shales and cherts with minor dark limestone beds and a few relatively thick-bedded debris-flow carbonates. Graptolites are common to abundant in the shales and conodonts occur in some of the thin, dark limestone beds. The Peel River section is a more or less typical Richardson Mountains boundary sequence, but is without significant carbonates.

The Road River strata of the Ogilvie Mountains, of which the Pat Lake and Blackstone River sections are representative, are characterized mainly by thinly bedded dark shales and calcareous shales, much greater amounts of dark limestone beds and laminae, and much less chert. The shales and calcareous shales contain abundant graptolites, while the dark limestones

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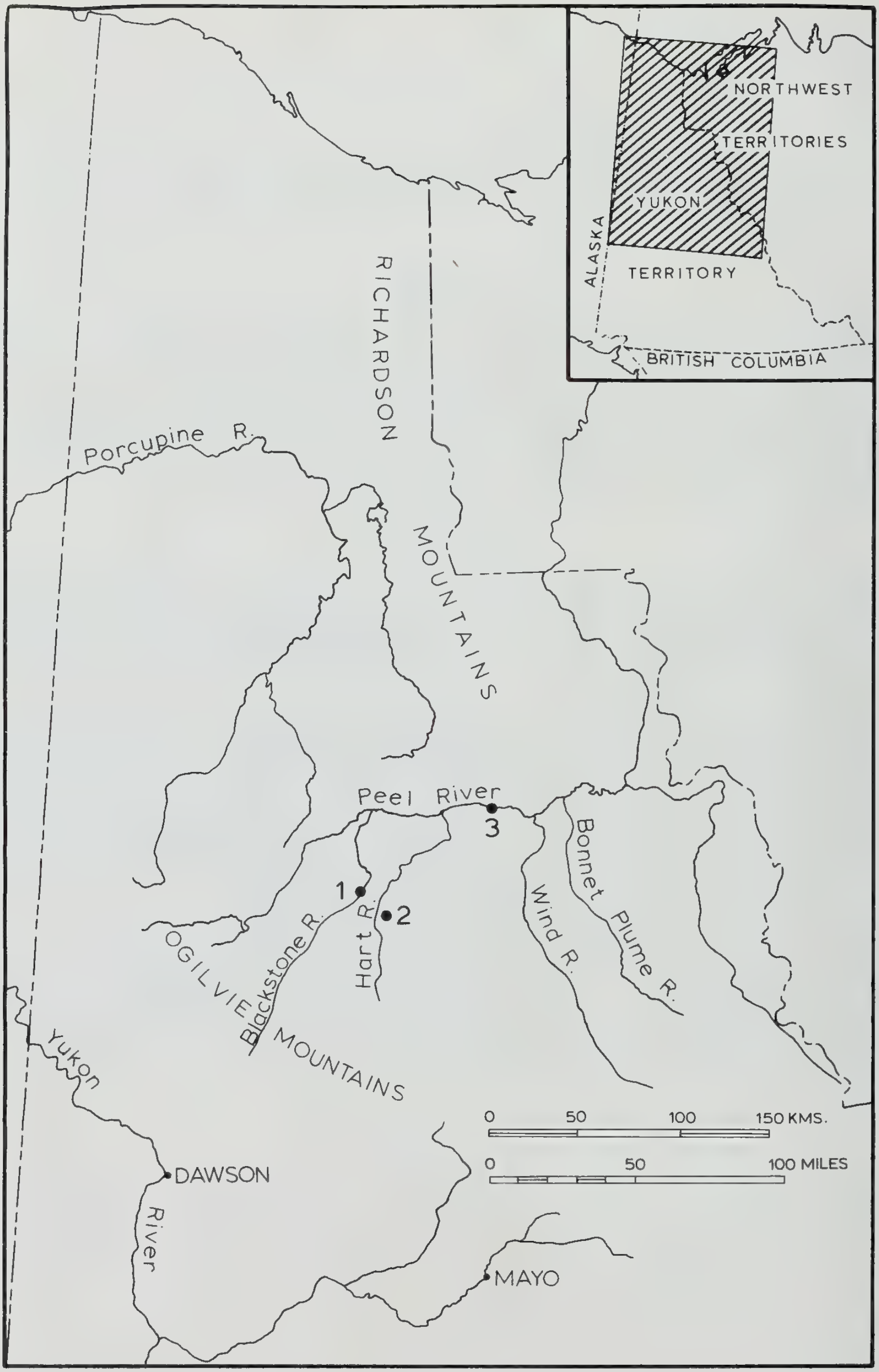
NORTHWEST
TERRITORIES
YUKON
ALASKA
TERRITORY
BRITISH COLUMBIA
RICHARDSON
MOUNTAINS
Porcupine R.
Peel River
3
1
2
Blackstone R.
Hart R.
Wind R.
Bonnet Plume R.
OGILVIE MOUNTAINS
Yukon
River
DAWSON
MAYO
0 50 100 150 KMS.
0 50 100 MILES

Fig. 1 Index map of northern Yukon showing localities. 1 = Blackstone River; 2 = Pat Lake; 3 = Peel River.

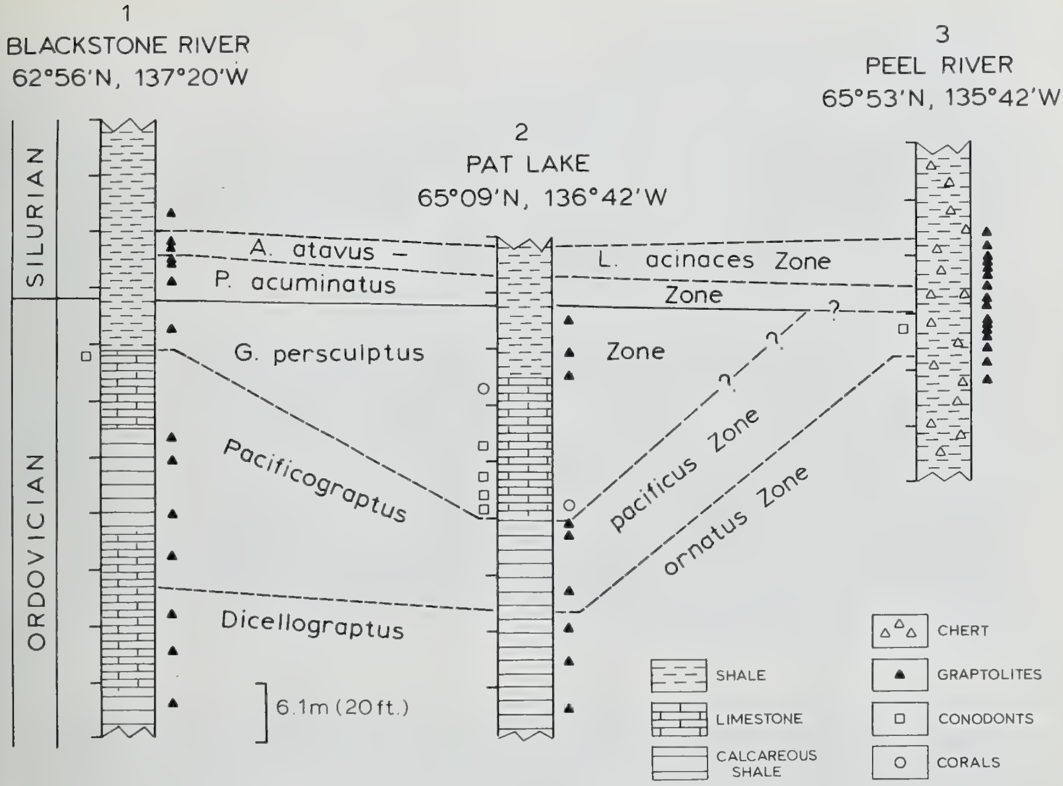
1
BLACKSTONE RIVER
62°56'N, 137°20'W
2
PAT LAKE
65°09'N, 136°42'W
3
PEEL RIVER
65°53'N, 135°42'W
SILURIAN
ORDOVICIAN
A. atavus –
P. acuminatus
Zone
L. acinaces Zone
G. persculptus
Zone
Pacificograptus
pacificus Zone
ornatus Zone
Dicellograptus
6.1m (20 ft.)
?
SHALE
LIMESTONE
CALCAREOUS SHALE
CHERT
GRAPTOLITES
CONODONTS
CORALS

Fig. 2 Correlation of graptolite zones of localites 1–3 (Fig. 1), using the base of the Silurian as a datum.

may contain conodonts and, rarely, trilobites (e.g. Lenz & Churkin 1966; Ludvigsen 1981). A relatively thick sequence of light-coloured, probably shallow water, conodont and coral-bearing limestone of probable latest Ordovician age (*Glyptograptus persculptus* Zone?) occurs in the Pat Lake section (Fig. 1). The presence of the limestone is anomalous, and its origin may be related to the widely recognized latest Ordovician glacially induced regression (e.g. Lenz 1976, 1982; Lenz & McCracken 1982).

Graptolites

Ashgill graptolite faunas of the northern Cordillera are divisible into two biostratigraphical units, a lower *Dicellograptus ornatus* Zone and the upper *Paraorthograptus pacificus* Zone. The uppermost Ordovician, the *G. persculptus* Zone, is less well developed and is clearly absent from the Peel River section, but is tentatively recognized in the Pat Lake and Blackstone River sections. Lowest Silurian (Llandovery) strata, represented by the *Parakidograptus acuminatus* Zone and the overlying *Atavograptus atavus* and *Lagarograptus acinaces* Zones are widely recognized (Lenz 1982).

The *D. ornatus* Zone is characterized by the index species, and by *D. minor*, *Glyptograptus latus*, *Climacograptus longispinus*, *C. latus*, *C. hvalross*, *C. hastatus*, *C. supernus*, *Orthograptus abbreviatus*, *O.* cf. *fastigatus*, *Orthoretiograptus denticulatus*, *Arachniograptus laqueus* and *Leptograptus* spp. *Dicellograptus* is common, as are most of the diplograptid species.

The *P. pacificus* Zone is taxonomically a much more impoverished fauna and is characterized by an abundance of *C. supernus* and *P. pacificus*. Most of the species of diplograptids noted in the *D. ornatus* Zone are present, but in much lesser numbers, and dicellograptids are rare. In addition, the exotic *Diceratograptus* cf. *mirus* is represented by two specimens in the Peel River section (Chen & Lenz 1984).

The supposed *G. persculptus* Zone, which was considered to be lowest Silurian in Lenz & McCracken (1982), is characterized by a fauna of low diversity, and is only tentatively recognized. The index species has not, to date, been recovered from the northern Canadian Cordillera, although it does occur in southeastern Alaska (Churkin *et al.* 1971). This biostratigraphical unit is distinguished by the relatively sudden appearance of narrow forms of *Climacograptus normalis* and *C. miserablilis*, a very spinose form of ?*Paraorthograptus* and *Orthograptus* cf. *abbreviatus*. Other species appearing in the interval, but not confined to it, include *Diplograptus modestus*, *Glyptograptus tamariscus*, *G. gnomus*, *G.* cf. *laciniosus*, and *G.* cf. *lanpheri*. Monograptids have not been recovered. The *G. persculptus* Zone? is absent in the Peel River section.

The *P. acuminatus* Zone, the lowest Silurian biostratigraphical unit, is readily recognized by the appearance of the index species, as well as *Climacograptus* cf. *trifilis*, ?*Akidograptus ascensus*, *Cystograptus vesiculosus* and *Diplograptus modestus diminutus*. Monograptids have not been found. The *A. atavus* and *L. acinaces* Zones are discussed together since they witness the incoming of monograptids, particularly *Atavograptus* and *Pribylograptus*, as well as being characterized by *Dimorphograptus confertus swanstoni*, *D. physophora* (and subspecies) and common *Cystograptus vesiculosus*.

Graptolite correlation

The graptolitic sequences of the northern Cordillera are directly comparable to those in central China and the Kolyma and Kazakhstan regions of the U.S.S.R., and indirectly with that of southern Scotland (Lenz & McCracken 1982; Chen & Lenz 1984). The *D. ornatus* Zone is directly comparable to the *C. longispinus* Subzone of Koren *et al.* (1979), more or less comparable with the *D. szechuanensis* Zone and possibly the *Amplexograptus yangtzeensis* Zone of central China (Chen & Lenz 1984), and probably with the *D. complanatus* Zone of Scotland (Williams 1982).

Correlation of the *P. pacificus* Zone of Yukon is almost certainly directly with the *P. pacificus* Subzone of U.S.S.R., but comparison with the Chinese succession is more difficult. Faunally, the *P. pacificus* Zone is most similar to the *D. szechuanensis* and *A. typicus* Zones; however, the presence of rare *Diceratograptus* in the Peel River section suggests correlation with strata as high as the *Paraorthograptus uniformis* Zone of China. The latter correlation would appear to be even more reasonable if *P. uniformis* of China is, as suggested by Williams (1982), synonymous with *P. pacificus*.

Correlation of the *G. persculptus* and *P. acuminatus* Zones is relatively straightforward, and it therefore appears that strata equivalent to the *Diplograptus bohemicus* Zone of China, and the *Climacograptus*? *extraordinarius* Zone of U.S.S.R. and Scotland are unrepresented by graptolites or missing from the Yukon sections.

Conodonts and conodont correlation

Ashgill conodonts from the Blackstone and Peel River sections are regarded as being within the *Amorphognathus ordovicicus* Biozone and the North American Fauna 12. The conodont fauna at Blackstone River (Figs 1, 2) occurs 3 m below the supposed *G. persculptus* Zone and 13·7 m above the last occurrence of graptolites of the *P. pacificus* Zone.

Significant taxa include *A. ordovicicus* Branson & Mehl, *Belodina confluens* Sweet, *Besselodus* n. sp., *Gamachignathus ensifer* McCracken *et al.*, *Icriodella superba* Rhodes?, *Noixodontus girardeauensis* (Satterfield), *Oulodus ulrichi* (Stone & Furnish), *Panderodus*? *sibber* Nowlan & Barnes, *Plectodina florida* Sweet, *P. tenuis* (Branson & Mehl), *Protopanderodus* sp., *Scabbardella altipes* (Henningsmoen) and *Walliserodus amplissimus* (Serpagli). Not all of these species were initially listed by Lenz & McCracken (1982) and some have since undergone taxonomic revision.

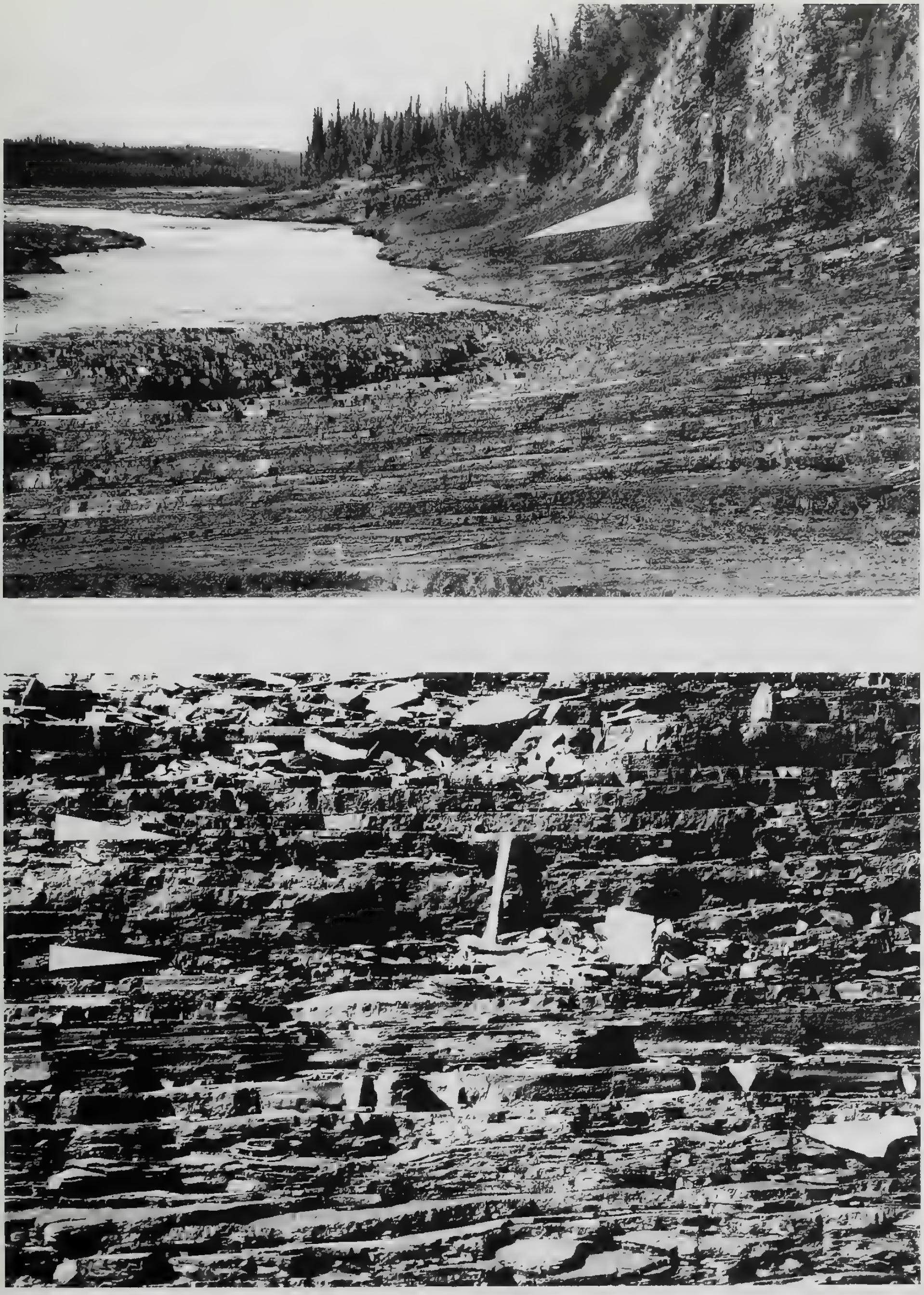

Fig. 3 Ordovician–Silurian boundary section on Peel River. Arrow on upper photograph is Ordovician–Silurian boundary. Lower photograph is a close-up of the boundary beds; the lower arrow is the top of the *P. pacificus* Zone and the upper arrow is the base of the *P. acuminatus* Zone.

One of the most noteworthy species, *N. girardeauensis*, was also found by McCracken & Barnes (1982) in Missouri in association with *Aphelognathus grandis* (Branson, Mehl & Branson) and *A. ordovicicus*. The recent work of Sweet (1984) established the *A. grandis* Chronozone; the nominal species not only occurs in the Missouri fauna, but also in the Richmondian Vauréal Formation of Anticosti Island (Nowlan & Barnes 1981). This species was not recognized in the Gamachian Fauna 13 by McCracken & Barnes (1981), but they recorded the related species *A.* aff. *A. grandis*. The range of *A. grandis* is reported to be from the upper Maysvillian through much of the Richmondian *A. divergens* Chronozone; it does not appear to range into post-Richmondian, pre-Silurian strata (Sweet 1984).

The close stratigraphical proximity of the Blackstone conodont fauna to the *G. persculptus* Zone? graptolites does not necessarily imply that it is latest Ordovician. The rare co-occurrence on Blackstone River of *G. ensifer* with *A. ordovicicus*, *B. confluens* (= *B. compressa* of Lenz & McCracken 1982), *O. ulrichi*, *P.? gibber*, *P. florida* and *P. tenuis* is comparable to the upper Vauréal Formation fauna (late Richmondian) of Nowlan & Barnes (1981). Unless the upper limit of *A. grandis* is younger than is at present known, the co-occurrence of *N. girardeauensis* and *A. grandis* in Missouri may indicate the Richmondian or, possibly, the late Maysvillian (based strictly on published microfossil data). Hence, the occurrence of *N. girardeauensis* at Blackstone River may favour a late, rather than the latest, Ordovician age. The Lower Llandovery shale and chert from both the Blackstone and Peel River sections have not been collected for conodonts.

A single *f* element of *G. ensifer* co-occurs at the Peel River section (Figs 1, 2) with some of the species listed above for the Blackstone River; *I. superba?*, *N. girardeauensis*, *O. ulrichi* and *P. florida* are absent from this fauna, whereas *O. rohneri* Ethington & Furnish and *Pseudobelodina vulgaris vulgaris* Sweet are present only at the Peel River section. Unlike the Blackstone section, where the Ordovician conodont fauna is within a thick, 16·7 m interval barren of graptolites, the Peel River conodont-bearing stratum is within a thin, 2·5 m interval bounded by shales containing graptolites of the *P. pacificus* Zone, and hence this conodont fauna is regarded as late, but not latest, Ordovician. The fauna occurs in strata 1·6–1·9 m below the systemic boundary.

Lenz & McCracken (1982) did not report Ashgill conodonts from the Pat Lake section Figs 1, 2). The sparse faunas there comprise poorly preserved conodonts that were originally assigned an early Silurian age on the basis of ramiform elements and on their stratigraphical

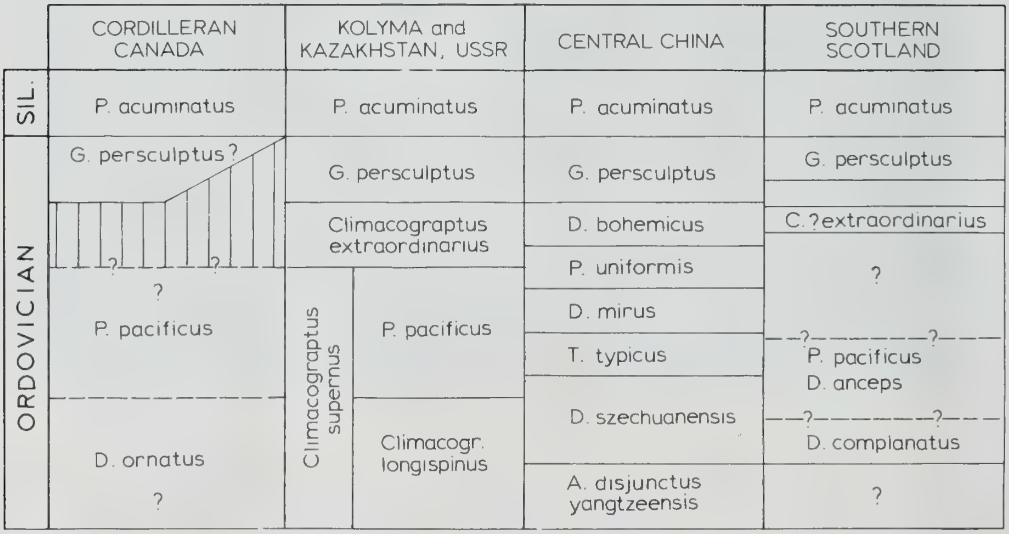
CORDILLERAN CANADA
KOLYMA and KAZAKHSTAN, USSR
CENTRAL CHINA
SOUTHERN SCOTLAND
SIL.
ORDOVICIAN
P. acuminatus
P. acuminatus
P. acuminatus
P. acuminatus
G. persculptus?
G. persculptus
G. persculptus
G. persculptus
?
?
Climacograptus extraordinarius
D. bohemicus
C. ?extraordinarius
P. uniformis
?
?
P. pacificus
Climacograptus supernus
P. pacificus
D. mirus
T. typicus
?
?
P. pacificus
D. anceps
D. szechuanensis
?
?
D. complanatus
D. ornatus
?
Climacogr. longispinus
A. disjunctus yangtzeensis
?

Fig. 4 Correlation of Ordovician–Silurian strata of Yukon with those of central China, U.S.S.R. and Scotland.

position with respect to *G. persculptus* Zone? graptolites. *Ozarkodina* sp. A Lenz & McCracken has a definite Silurian aspect, although this does not demand an assignment to that system since the genus has elsewhere been occasionally recognized from Upper Ordovician strata.

The poor preservation of the coniform elements limits their biostratigraphical value; they could be assigned to either Ordovician or Silurian taxa. Thus an age determination for these post-*P. pacificus* Zone and pre-*persculptus* Zone? conodonts may depend upon the positive identification of *Ozarkodina* sp. A. An unequivocal age based solely on the conodont taxa cannot be determined for the Pat Lake conodont faunas. Their occurrence below the *G. persculptus* Zone? suggests an Ordovician age.

Llandovery and younger conodont faunas are much more diverse and better preserved than those discussed herein; study of these faunas is in progress.

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The Ordovician–Silurian boundary in the United States

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Synopsis

Ordovician and Silurian rocks are widespread in the United States and there are numerous outcrops in many regions displaying the systemic boundary interval. However, a regional review of key sections in all the major outcrop areas shows that biostratigraphically closely controlled and stratigraphically complete or nearly complete boundary successions are quite rare. Indeed, the Esquibel Island section in south-eastern Alaska, where the systemic boundary is in a continuous graptolitiferous sequence, is not only the only known occurrence in the United States of a typical *P. acuminatus* Zone fauna, but also the only known place in the country where the systemic boundary can be established precisely on graptolites in a continuous succession. Elsewhere, relatively complete, if not complete, boundary successions are present in the Appalachians and in the Great Basin, as well as Alaska, but in virtually all cases the biostratigraphical control is not good enough to establish the boundary level with certainty. Most of the sections in these regions display a gap in the boundary interval, and this is the case also in most of the many boundary sections in the Midcontinent region. The best known, and stratigraphically most nearly complete, cratonic sections are in Arkansas, Oklahoma, Missouri, and Illinois, where strata having a taxonomically varied *Hirnantia* fauna are overlain, with locally only a minor, if any, stratigraphical gap, by rocks containing Llandovery fossils. No graptoloid graptolites are known from these sections, and the precise level of the systemic boundary is uncertain in some sections. It is concluded that further studies are urgently needed on fossils and rocks in the boundary interval, particularly to establish the precise age of the conodont faunal turnover as well as to clarify the mutual relations between the distribution patterns in time and space displayed by different groups, and their relations to the graptolite-based systemic boundary.

Introduction

Ordovician rocks are present in the subsurface over much of the United States and they are exposed in several major regions (Cook & Bally 1975). Although less widespread than those of Ordovician age, Silurian rocks are likewise distributed over major parts of the country and exposed over considerable areas. Accordingly, it is not surprising that the interval of the Ordovician–Silurian systemic boundary is available for study at a large number of localities from the Appalachians in the east to the Great Basin in the west. In many of these sections, the faunal succession is incompletely known or fossils are absent in critical intervals, which applies to the cratonic areas in the continental interior as well as to the geosynclinal areas along the continental margins. Nevertheless, because in most sections, particularly the cratonic ones, the systemic boundary is associated with a stratigraphical gap and a change in lithology, its level in those sections can be readily recognized. As is the case elsewhere in the world, nearly complete successions in continuously fossiliferous facies across the boundary interval are quite rare in the United States both in shelly and graptolitic facies. For instance, we are not aware of a single section outside Alaska where the precise level of the base of the *P. acuminatus* Zone, that is the internationally accepted base of the Silurian, can be recognized by means of graptolites or other fossils. It is quite clear that the choice of this level for the systemic boundary at the present time makes its recognition difficult, if not impossible, in stratigraphically more or less complete successions like those in the Great Basin (Ross *et al.* 1979; Leatham 1985; etc.) and in the Mississippi Valley region (Amsden 1986).

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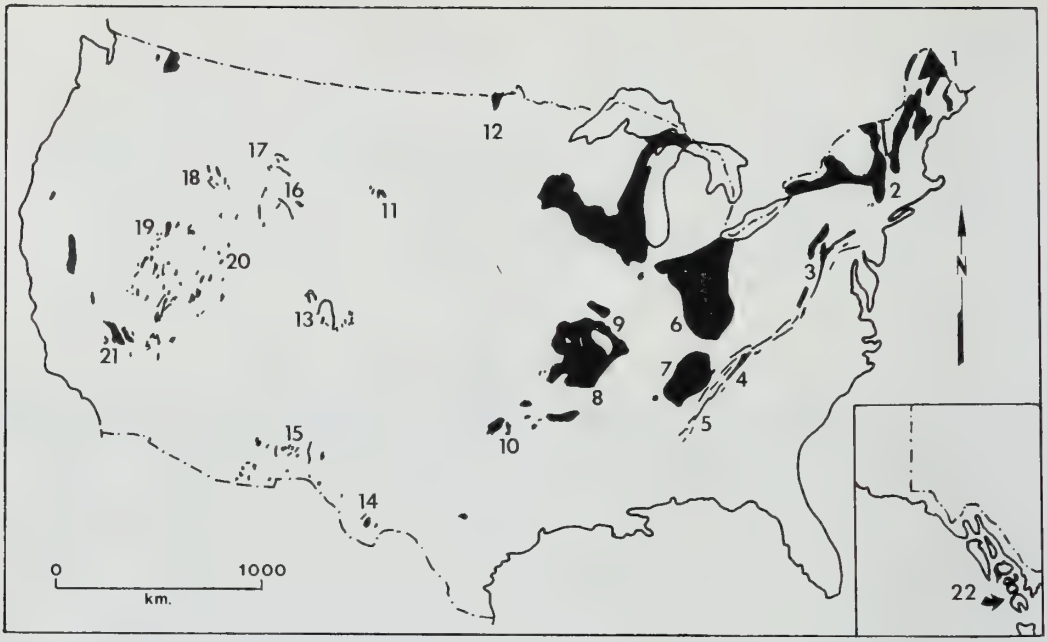

Fig. 1 Index map showing areas with Ordovician and/or Silurian outcrops (black) and systemic boundary sections. 1, northern Maine; 2, eastern New York and western Vermont; 3, central Appalachians (Pennsylvania and adjacent states); 4, eastern Tennessee; 5, Alabama and Georgia (southern Appalachians); 6, the Cincinnati region and adjacent areas in Ohio, Kentucky, and Indiana; 7, the Nashville dome in central Tennessee; 8, northern Arkansas (including the Batesville district); 9, eastern Missouri and southwestern Illinois; 10, southern Oklahoma (including the Arbuckle Mountains); 11, Black Hills (South Dakota and Wyoming); 12, North Dakota; 13, Colorado; 14, west Texas; 15, New Mexico; 16, Bighorn Mountains, Wyoming; 17, Montana; 18, Idaho; 19, Nevada; 20, Utah; 21, southeastern California; 22, southeastern Alaska (inset map).

The purpose of the present paper is to review briefly the biostratigraphy of the systemic boundary interval in key sections in the principal outcrop areas. Page limitations make it necessary to restrict ourselves to data essential for the understanding of the local and regional geology of this interval in the United States. For convenience, we will deal with each of the major outcrop regions separately, from the Appalachians in the east to the Great Basin in the west. For the location of these regions, see Fig. 1.

Northern Appalachians

In large parts of the Northern Appalachians in the United States (Maine to New York State), Silurian or younger rocks rest with a conspicuous, in many cases angular, unconformity on the Ordovician (Berry & Boucot 1970: fig. 6). This stratigraphical gap varies in magnitude both locally and regionally but includes in most cases portions of both the Ordovician and Silurian systems. Conventionally, this gap is explained as a product of the Middle to Late Ordovician Taconic orogeny, but it is evident that the apparently global drop in sea level during the latest Ordovician (Hirnantian) contributed to emergent conditions, at least locally.

In this region, biostratigraphical control through the systemic boundary interval is, in general, poor. This is partly due to the fact that the rocks were largely deposited in environments with small numbers of shelly organisms, and those that became fossilized were in many cases strongly affected by the subsequent metamorphism of the host rocks. That diagnostic fossils are present locally is shown by Neuman's (1968) finds of shelly fossils of the *Hirnantia* fauna in east-central Maine, the only occurrences of this type of fauna from the northeastern United States. Another, and in terms of geology of the systemic boundary even more interesting, sequence is that of the Carys Mills Formation of northeastern Maine and adjacent New Brunswick. The lower part of this thick unit has yielded specimens of *Glyptograptus persculptus* (Rickards & Riva 1981) and Llandovery age graptolites are known from higher parts of the formation (Pavlides 1968). The Carys Mills has also produced well preserved conodonts of the *Icriodella discreta–I. deflecta* Zone of probable Rhuddanian (early Llandovery) age (Barnes & Bergström, this volume) but, unfortunately, the precise stratigraphical position of these conodonts within the formation is uncertain because of the scattered exposures, considerable thickness, monotonous lithology, and structural deformation of the unit. At any rate, it appears rather likely that the Carys Mills represents a stratigraphically complete succession from the uppermost Ordovician to the lower Silurian, but further studies are needed to pinpoint the level of the systemic boundary.

Central and Southern Appalachians

In southern New York and parts of eastern Pennsylvania and Virginia (Fig. 1), the Ordovician–Silurian boundary is marked by an unconformity (Dennison 1976) and parts of the Ordovician, and possibly also of the lowermost Silurian, are missing. From north-central Pennsylvania to eastern Tennessee, the systemic boundary is somewhere in a succession, several hundred metres thick, of near-shore to non-marine clastic sediments lacking shelly fossils of stratigraphical utility. Although the precise level of the systemic boundary remains undetermined in these successions, it has been common practice to classify the Juniata and Sequatchie formations as Ordovician and the overlying Tuscarora and Clinch formations as Silurian.

Recent work by Colbath (1986) has raised the possibility of establishing a viable palynomorph (acritarch and chitinozoan) biostratigraphy useful for precise recognition and correlation of the systemic boundary in these successions. Likewise, Gray's work (1985) on higher land plant spore tetrads permits recognition of the approximate boundary interval. Both the spore tetrads and the marine palynomorphs occur in some abundance in near-shore marine sediments. The spores are also found in purely non-marine facies provided they have not been destroyed by low-temperature metamorphism of the host strata. However, palynomorph work in the systemic boundary interval in this region has not passed the pioneer stage, and much additional study is needed to assess the local and regional biostratigraphic utility of these fossils.

In the southernmost Appalachians, in the Birmingham area of Alabama, the systemic boundary is marked by a conspicuous stratigraphical gap that includes the entire Upper Ordovician and probably the lowermost Llandovery as well (Hall, unpublished; Berry & Boucot 1970). Near the Alabama–Georgia boundary, the stratigraphical gap also includes the entire Middle Ordovician (Dennison 1976), but in northwesternmost Georgia, Chowns (1972) considered the systemic contact conformable on lithological evidence. The youngest Ordovician strata in much of Alabama, which are referred to the Sequatchie Formation (Drahovzal & Neathery 1971), are of Late Ordovician (Maysvillian and Richmondian) age. In Limestone County in northern Alabama, the Devonian Chattanooga Shale contains reworked Late Ordovician (probably Richmondian) conodonts (Bergström, unpublished) apparently originating from now-eroded rocks that may be younger than the biostratigraphically dated parts of the Sequatchie Formation. Where dated biostratigraphically, the Sequatchie is separated from overlying rocks by a stratigraphical gap corresponding not only to the uppermost Ordovician but also some part of the post-Ordovician succession. Locally this gap is substantial and may include more than a system.

Eastern North American Midcontinent

We include in this area the Cincinnati Arch region in Ohio, Kentucky, and Indiana, and the Nashville Dome area in central Tennessee (Fig. 1).

The Cincinnati region contains the Reference Standard of the North American Upper Ordovician, the Cincinnatian Series. Both faunal and lithological evidence suggest an appreciable hiatus between the Ordovician and the Silurian over the entire outcrop area in the Cincinnati region. The stratigraphically most complete succession is apparently on the eastern side of the Cincinnati Arch in southern Ohio and adjacent Kentucky. There is no record of Hirnantian (latest Ordovician) age rocks anywhere in the Cincinnati region and the youngest Cincinnatian stage, the Richmondian, is considered to be of pre-Hirnantian age. Based on the succession of Anticosti Island, Québec, Canada, Twenhofel's (1921) Gamachian Stage has in recent years been recognized as a post-Richmondian, pre-Silurian standard unit (Barnes & McCracken 1981). Although rocks of Gamachian age are not known to be represented in the Cincinnatian type area, the Gamachian is now classified as the uppermost part of the Cincinnatian Series (Ross *et al.* 1982).

One of the best exposed and most representative sections through the Ordovician–Silurian boundary interval on the eastern flank of the Cincinnati Arch is a series of exposures along Ohio Highway 41 between West Union and Ohio Brush Creek, Adams County, Ohio (Summerson 1963; Rexroad *et al.* 1965; Gray & Boucot 1972; Grahn & Bergström 1985). In this section, the beds are horizontal, developed in fossiliferous limestone and shale, and there are no structural complications. The topmost Ordovician unit, the Drakes Formation of Richmondian age, is overlain conformably and without conspicuous lithological break by the Belfast Member of the Brassfield Formation (Fig. 2). This unit has produced a relatively undiagnostic conodont fauna of general early Llandovery type (Rexroad 1967) as well as chitinozoans suggesting a *C. cyphus* Zone age (Grahn & Bergström 1985). Grahn & Bergström (1985) interpreted the stratigraphical gap as corresponding to about four graptolite zones and it is surprising that there is no channelling, development of a conglomerate, or other lithic evidence of a sedimentary break. The major body of the Brassfield, that is, its post-Belfast part, contains a rich megafossil fauna of early to middle Llandovery age (Berry & Boucot 1970) as well as a stratigraphically diagnostic conodont fauna of the *Distomodus kentuckyensis* Zone (Rexroad 1967; Cooper 1975) and chitinozoans (Grahn 1985). There are no graptolites known from this succession.

In many other Cincinnati region sections, especially on the west flank of the Cincinnati Arch, the stratigraphical gap associated with the systemic boundary is even greater than in the Ohio Brush Creek sections (Rexroad & Kleffner 1984).

In parts of the Nashville Dome in central Tennessee, the Devonian Chattanooga Shale unconformably overlies Middle Ordovician rocks (Dennison 1976). In other parts of the Nashville Dome, strata dated as Richmondian are overlain unconformably by the Brassfield Limestone of middle Llandovery age (Wilson 1949), which indicates the presence of a stratigraphical gap of magnitude similar to that in the Cincinnati region.

Central North American Midcontinent

We include in this area Oklahoma and adjacent Texas Panhandle, Arkansas, Missouri, Illinois, Minnesota, and Wisconsin (Fig. 1).

In a recent comprehensive study, Amsden & Barrick (1986) provided a useful summary of the geology of the Ordovician–Silurian boundary interval in this region. Of particular significance is the confirmation of the widespread occurrence of latest Ordovician strata having shelly fossils of the *Hirnantia* fauna and conodonts of the *Noixodontus* fauna. The stratigraphically most informative sections are in the Batesville district of north-central Arkansas and in eastern Missouri. Both locally and regionally, the stratigraphical succession varies a great deal, and in several cases, sections in close proximity to each other exhibit striking differences in lithological and stratigraphical development. This is well illustrated by the conditions in the Batesville district as well as in eastern Missouri.

In the Batesville district two sections are of particular interest. One of these sections is in the Love Hollow Quarry (Craig 1968, 1986*a*, 1986*b*; Amsden 1968, 1986). In this large and recently active quarry, the beds are horizontal and there are no notable tectonic complications. A

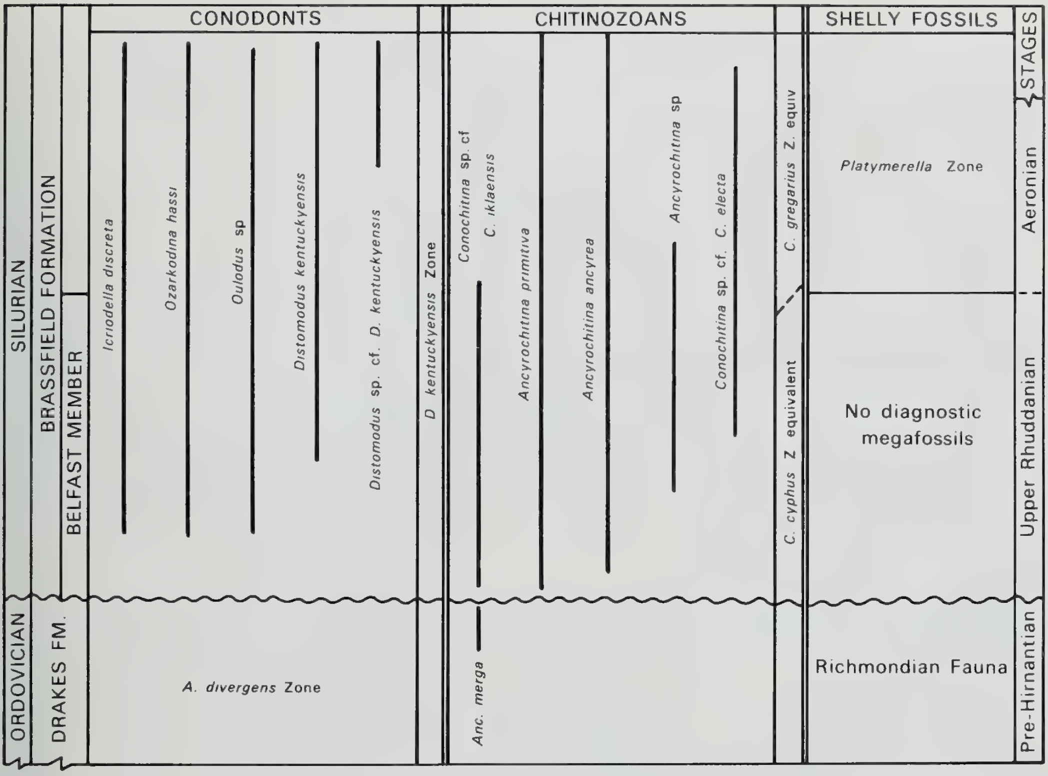

Fig. 2 Vertical ranges of selected conodont and chitinozoan species, and the occurrence of index megafossil assemblages, in the systemic boundary interval in exposures along Ohio Highway 41 northeast of West Union, Adams County, Ohio. Based on data from Berry & Boucot (1970), Cooper (1975), Grahn & Bergström (1985), and Grahn (1985). Note that there is a prominent stratigraphical gap between the Ordovician and the Silurian corresponding to the Hirnantian and the lower Rhuddanian. Although this gap is about four graptolite zones, there is very little lithological evidence of its existence in these sections.

stratigraphical column with fossil occurrences is given in Fig. 3. It should be noted that the Cason Oolite as well as the overlying *Triplesia alata* beds were developed in a large limestone lens which is now quarried away.

The Cason Oolite contains brachiopods that are used by Amsden (1986) for correlation with the Hirnantian Keel Limestone of Oklahoma. The oolite also contains conodonts of the *Noixodontus* fauna (Craig 1986*a*; Barrick 1986) that supports this correlation. The overlying pelmatozoan limestone, referred to by Amsden (1986) as the *Triplesia alata* beds and by Craig (1986*b*) as the Brassfield Limestone, contains late Llandovery brachiopods and conodonts (Craig 1986*b*). No graptolites have been found in this succession. The contact between the Cason Oolite and the overlying pelmatozoan limestone has been described as 'stylolitic' (Craig 1969). It appears to represent a stratigraphical gap but its exact magnitude is uncertain, although Barrick (1986) and Craig (1986*b*) report *O. celloni* Zone (late Llandovery) conodonts from the *Triplesia alata* beds at this locality.

A similar succession (Fig. 3) is reasonably well exposed 0·5 km NE of St. Clair Springs (Amsden 1986) in which the Cason Oolite, which contains a Hirnantian age brachiopod fauna similar to that of the Keel of Oklahoma and the Edgewood of Missouri, is directly overlain by about 3 m of crinoidal limestone classified as the Brassfield Limestone by Craig (1986*b*). The

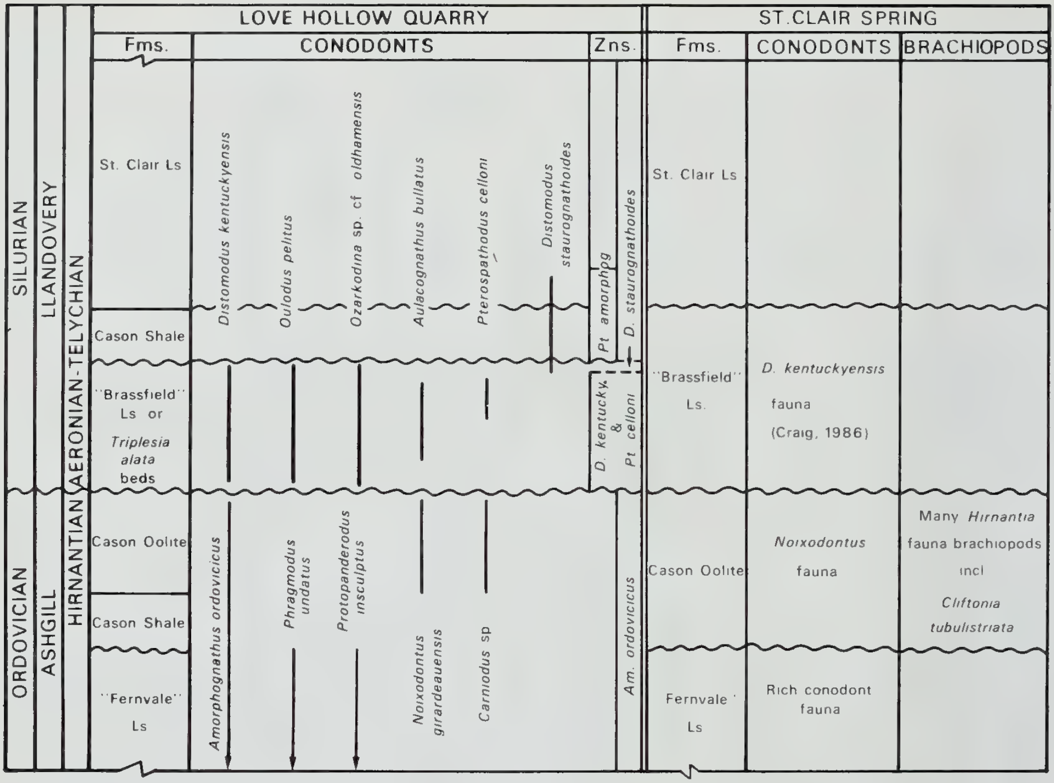

Fig. 3 Vertical ranges of important conodont species, and the occurrence of *Hirnantia* fauna brachiopods in two sections in the Batesville district, Arkansas. Based on Amsden (1986), Craig (1986*a*, 1986*b*), and Barrick (1986). Note that there is a conspicuous stratigraphical gap in the systemic boundary interval with a considerable portion of the Llandovery missing. The Love Hollow Quarry exposure of the Cason Oolite and the *Triplesia alata* beds is now quarried away (Amsden 1986).

conodonts from this locality confirm that the Cason Oolite is of Hirnantian age and that the overlying Brassfield is coeval with the Brassfield of the Cincinnati region (Craig 1986*b*; Barrick 1986). The systemic boundary is placed at the base of the Brassfield and is not strongly expressed lithologically; it may be associated with a stratigraphical gap corresponding to the lowermost Llandovery, but conodonts and other fossils do not provide sufficient stratigraphical resolution to assess its magnitude precisely.

The Cason Oolite equivalent in southern Oklahoma is apparently the Keel Limestone (Amsden 1986) that has yielded Hirnantian age brachiopods as well as conodonts of the *Noixodontus* fauna (Barrick 1986). Its topmost part has also produced stratigraphically younger conodonts of general Silurian aspect but no Silurian index species. Barrick assigned the latter fauna to the Llandovery and placed the systemic boundary within the Keel. Amsden, on the basis of his brachiopod studies, placed the entire Keel in the Ordovician (Amsden 1986: text-fig. 37) and noted that the unit is separated from the overlying Cochrane Formation by a large stratigraphical gap corresponding to the lower and middle Llandovery. In our opinion, the Silurian-type conodont fauna reported from the upper Keel by Barrick (1986) does not provide firm evidence of Silurian age because, as shown by Barnes & Bergström (this volume, p. 325), the turnover from an Ordovician-type to a Silurian-type conodont fauna may well have taken place in very latest Ordovician (late *G. persculptus* Zone) time, within a time interval older than the base of the Silurian. Whether or not this alternative dating is correct can be solved only after the conodont faunal turnover has been firmly dated in terms of graptolite zones.

As noted by Amsden (1986), there are two important outcrop areas of the systemic interval in the Mississippi Valley, one in west-central Illinois and northeastern Missouri, and the other in southwestern Illinois and southeastern Missouri. A considerable number of sections through the uppermost Ordovician and overlying Silurian strata have been described by Amsden (1974, 1986) and Thompson & Satterfield (1975). The former also described the brachiopod faunas and the latter reported on the conodonts (also cf. McCracken & Barnes 1982). The stratigraphically most complete systemic boundary sequences are in the former area; in the latter area, the Edgewood Group, of Hirnantian age at the top, is overlain directly and unconformably by the Sexton Creek Limestone that contains brachiopods suggesting a late Llandovery (late Aeronian–Telychian) age (see Fig. 4).

One of the biostratigraphically most instructive sections is along the west side of Missouri Highway 79 at Clinton Springs at the south edge of Louisiana, Pike County, Missouri, where the horizontal beds are easily accessible along a major highway. Good brachiopod collections of Hirnantian age have been described from the Noix Oolite at this locality (Amsden 1974, 1986) and conodonts (of *Noixodontus* fauna type) studied by Thompson & Satterfield (1975). The overlying Bryant Knob Formation has yielded a few brachiopods (Amsden 1974) and conodonts interpreted as indicating early Llandovery age (Thompson & Satterfield 1975). The

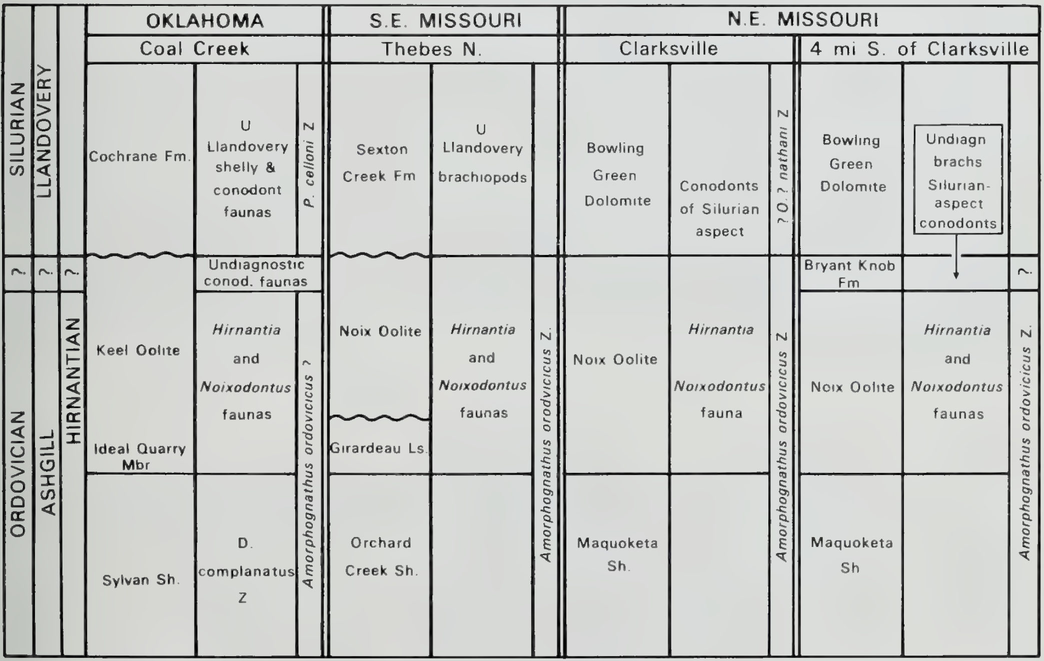
SILURIAN
LLANDOVERY
ORDOVICIAN
ASHGILL
HIRNANTIAN
?
OKLAHOMA
Coal Creek
Cochrane Fm.
U Llandovery shelly & conodont faunas
P. celloni Z
Undiagnostic conod. faunas
Keel Oolite
Ideal Quarry Mbr
Hirnantia and Noixodontus faunas
Amorphognathus ordovicicus ?
Sylvan Sh.
D. complanatus Z
S.E. MISSOURI
Thebes N.
Sexton Creek Fm
U Llandovery brachiopods
Noix Oolite
Girardeau Ls.
Hirnantia and Noixodontus faunas
Amorphognathus ordovicicus Z.
Orchard Creek Sh.
N.E. MISSOURI
Clarksville
Bowling Green Dolomite
Conodonts of Silurian aspect
? O. ? nathani Z
Noix Oolite
Hirnantia Noixodontus fauna
Amorphognathus ordovicicus Z
Maquoketa Sh.
4 mi S. of Clarksville
Bowling Green Dolomite
Undiagn brachs Silurian-aspect conodonts
Bryant Knob Fm
Noix Oolite
Hirnantia and Noixodontus faunas
Amorphognathus ordovicicus Z.
Maquoketa Sh

Fig. 4 Occurrence of key fossil assemblages, and general biostratigraphy, in the systemic boundary interval at some localities in Oklahoma, southeastern Missouri, and northeastern Missouri. For the location of these sections, see Amsden & Barrick (1986), and Thompson & Satterfield (1975), and these papers provide most of the data upon which this diagram is based. As is clear from the diagram, it is a review of the general stratigraphy in each of the areas and no correlation is implied between a unit in one column and one at the same vertical position in another column. In Oklahoma and southeastern Missouri, the systemic boundary is associated with a prominent stratigraphical gap whereas in the illustrated sections from northeastern Missouri, the succession across the systemic boundary may have only a minor, if any, stratigraphical gap.

succession does not show any distinct lithic break between these units and it may be one of the stratigraphically most nearly complete boundary successions in the Midcontinent region. A stratigraphically similar section is present along Highway 79 about 6·5 km south of Clarksville and about 19 km southeast of the Clinton Springs locality (Fig. 4; Amsden 1974). In his recent reassessment of the data at hand, Amsden again placed the systemic boundary at the top of the Noix Oolite but indicated (1986: 42) that 'the brachiopod biostratigraphy requires no significant interruption in the Noix-Bryant Knob sequence'. Interestingly, McCracken & Barnes (1982) reported conodonts of Silurian aspect, by them interpreted as representing either the *O.*? *nathani* Zone or the *D. kentuckyensis* Zone, from the lowermost 1·65 m of the Bowling Green Dolomite from a locality near Clarksville, where this unit directly overlies the Noix Oolite, which yielded a representative *Noixodontus* fauna. The conodont faunas from the Noix and the Bowling Green are quite different, and there is obviously a faunal turnover between these units. Unfortunately, as noted by Barnes & Bergström (this volume), the precise age of this faunal turnover is currently unknown in terms of the graptolite succession, but it is quite possible that it took place in the latest Ordovician. If so, it cannot be excluded that the systemic boundary is above the base of the Bowling Green. However, the fact that the latter unit is overlain by the late Llandovery Sexton Creek Limestone (Amsden 1986) makes it clear that the systemic boundary must be below the base of the latter unit.

Western Midcontinent

Important outcrop areas in this vast region (Fig. 1) include the Black Hills in South Dakota, the Bighorn Mountains in Wyoming, and areas in Montana, Colorado, southern New Mexico, and western Texas. Most of the Upper Ordovician in these areas consists of shallow-water carbonates with few megafossils but with taxonomically varied and biostratigraphically useful conodont faunas (Sweet 1979). The biostratigraphy of the overlying beds is less well known. No biostratigraphically well-controlled section is currently known that is stratigraphically reasonably complete in the Ordovician–Silurian boundary interval, and the data suggest that everywhere rocks of Ordovician age are separated from younger rocks by an unconformity representing a significant stratigraphical gap (Ross *et al.* 1982). The most nearly complete boundary section may be in the subsurface of North Dakota; however, data from the depositional basin extension in adjacent Manitoba, where the succession is quite similar to that in North Dakota, suggest the absence of at least the lowermost Llandovery (Barnes & Bergström, this volume).

Great Basin

We include in this region western Utah, Nevada, Idaho, and southern California (Fig. 1). There are numerous excellent sections of Upper Ordovician and Lower Silurian rocks in carbonate facies with virtually 100% exposure in the Great Basin, and most of these sections may be reached by car and by foot under reasonable conditions. However, many localities are structurally complex, and widespread secondary dolomitization, particularly in the Ordovician, makes it difficult to obtain well-preserved megafossils. Furthermore, diagnostic shelly megafossils are not common and graptolites are rare. Conodonts are known from some sections and they offer great potential for detailed stratigraphical work in the widespread carbonates; unfortunately, the problem of dating the conodont faunal turnover referred to above currently restricts their use in establishing precisely the position of the Ordovician–Silurian boundary. Accordingly, it is currently impossible to recognize with certainty the exact level of the systemic boundary, or even to assess whether or not deposition was continuous, at those carbonate sections where there is not a conspicuous unconformity in the boundary interval.

Much of the pertinent biostratigraphical information from megafossils was summarized by Berry & Boucot (1970). Additional data from shelly fossils have been published by, among others, Budge & Sheehan (1980*a*, 1980*b*) and Sheehan (1980, 1982).

Although conodont work in the systemic boundary interval is still in the pioneer stage in the Great Basin, it is apparent that conodonts offer greater potential than any other group for detailed biostratigraphical work. Two recent conodont studies deserve mention in a discussion of the systemic boundary. Ross *et al.* (1979) described the conodont biostratigraphy of the Hanson Creek Formation near Eureka, Nevada. They suggested that this unit represents continuous deposition from Ordovician to Silurian time. This is quite possible, but it is perhaps equally possible that all the conodont samples referred to in their study are of Ordovician age and that the systemic boundary is at a higher, as yet undetermined, level in the Hanson Creek than that advocated by Ross *et al.* because, as noted by Barnes & Bergström (this volume), none of their conodont collections contain index conodonts of definite Silurian age.

In another recent study, Leatham (1985) described the conodont biostratigraphy of the Fish Haven Dolomite and immediately overlying strata in a section in northernmost Utah. He identified a prominent conodont faunal turnover and a transitional fauna interval of 5·5 m thickness in the uppermost Fish Haven. The systemic boundary was placed at the base of this transition interval, but Leatham (1985) was uncertain whether or not there was a stratigraphical gap at this level. He was also uncertain about the nature of the mixed faunal association and suggested that it might be a product of reworking or stratigraphic leak. In our view, it cannot be excluded that the interval with the mixed fauna, regardless of its nature, is of Hirnantian age, and that the systemic boundary, as it is now defined by means of graptolites, is at a somewhat higher stratigraphical level, in the lowermost part of the Laketown dolomite.

Of special interest in a review of the Ordovician–Silurian boundary biostratigraphy in the Great Basin is Berry's (1986) record of an uppermost Ordovician to lower Silurian sequence of graptolite faunas in cherts and dolomites of the upper Hanson Creek Formation in the Monitor Range, central Nevada. A quartz sand-bearing dolomite, which evidently represents a period of shallowing near the end of the Ordovician, is underlain by strata having the *Dicellograptus complanatus ornatus* graptolite assemblage, and directly overlain by rocks containing the diagnostic species association of the *Glyptograptus persculptus* Zone. Stratigraphically higher beds contain species that may represent the *P. acuminatus* Zone but the zonal index has not been found.

Graptolite-bearing shale sequences of Ordovician and Silurian age are widespread in the mountain ranges in the Great Basin but the studied successions appear to be stratigraphically incomplete and display a gap in the systemic boundary interval. For instance, in the carefully studied and well-known graptolite shale succession in the Trail Creek area, central Idaho, Llandovery beds older than the *M. convolutus* Zone are missing (Carter & Churkin 1977).

Alaska

With one important exception, little information is currently available concerning the geology of the Ordovician–Silurian boundary in Alaska. This exception is the Prince of Wales region in southeastern Alaska (Fig. 1) where in the long-ranging Descon Formation there is a quite condensed succession through the systemic boundary interval, which displays a complete sequence of late Ordovician–early Silurian graptolite zones. The best known succession is on Esquibel Island (Churkin & Carter 1970; Churkin *et al.* 1971) where a few metres thick sequence of cherty shales spans the systemic boundary without any indication of depositional breaks. A less than 3 m thick interval with the *G. persculptus* Zone fauna is overlain by about 1·5 m of strata containing graptolites characteristic of the *P. acuminatus* Zone, including the zonal index. The Esquibel Island graptolite species associations show close similarity to those of coeval strata in the Birkhill Shale in the Ordovician–Silurian boundary stratotype at Dob's Linn in south Scotland, making it possible to recognize the level of the systemic boundary with considerable precision. This may be the only place in the United States where the level of the systemic boundary can be fixed conclusively on graptolite evidence in a stratigraphically continuous section, and one can only regret that this key locality is located in a remote region that is likely to be visited by very few geologists.

Conclusions

1. The Ordovician–Silurian boundary interval is well exposed at numerous localities throughout the United States from the Appalachians in the east to the Great Basin in the west.
2. Available biostratigraphical and/or lithostratigraphical evidence suggests that in the vast majority of these sections, there is a stratigraphical gap, of greatly different magnitude in different sections, in the boundary interval (Fig. 5). Particularly in the shallow-water cratonic successions, this gap reflects the global drop of sea-level near the end of the Ordovician, but there is evidence that local uplifts have been of importance in some areas. Currently, we are aware of only a single biostratigraphically closely controlled section in the United States, on Esquibel Island, southeastern Alaska, which displays continuous deposition throughout the boundary interval. However, such sections may exist elsewhere, particularly in the Appalachians and in the Great Basin.
3. Some of the best, and biostratigraphically most closely controlled, boundary sections are in Arkansas, Oklahoma, Missouri, and Illinois where rocks having the *Hirnantia* shelly fauna and the *Noixodontus* conodont fauna are overlain by Llandovery-age strata with, at least locally, only a minor, if any, stratigraphical gap. Regrettably, no stratigraphically diagnostic graptolites are known from these sections.
4. A considerable number of well-exposed, thick, and apparently stratigraphically relatively complete sections in shallow-water carbonate facies are known from the Great Basin. Dolomitization has seriously affected the state of preservation of the megafossils, which are rather scarce in most sections, but conodonts are moderately common and taxonomically varied. Yet, because the conodont biostratigraphy is not tied reliably to the graptolite zone succession in the *G. persculptus* and *P. acuminatus* Zones, currently the conodonts cannot be used to pinpoint the level of the systemic boundary in carbonate sections without a significant stratigraphical gap.
5. As far as we are aware, in the United States the *P. acuminatus* Zone has been identified with certainty only on Esquibel Island, Alaska, and this is the only place where the level of the systemic boundary can be established precisely by means of zonal graptolites. The successions of shelly fossils, conodonts and palynomorphs are thus far calibrated only impre-

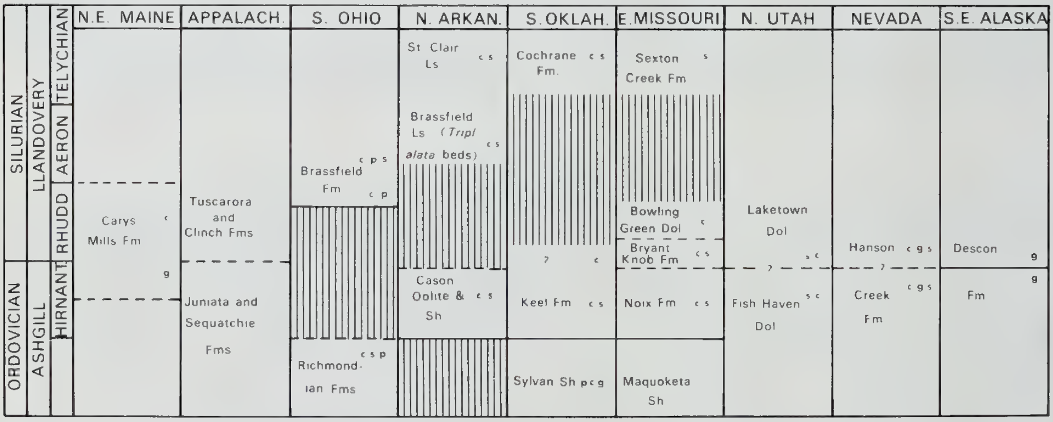
SILURIAN
ORDOVICIAN
LLANDOVERY
ASHGILL
TELYCHIAN
AERON
RHUDD
HIRNANT
N.E. MAINE
APPALACH.
S. OHIO
N. ARKAN.
S. OKLAH.
E. MISSOURI
N. UTAH
NEVADA
S.E. ALASKA
Carys Mills Fm c
g
Tuscarora and Clinch Fms
Juniata and Sequatchie Fms
Brassfield Fm c p s c p
Richmondian Fms c s p
St Clair Ls c s
Brassfield Ls (Tripl alata beds) c s
Cason Oolite & Sh c s
Cochrane Fm. c s
? c
Keel Fm c s
Sylvan Sh p c g
Sexton Creek Fm s
Bowling Green Dol c
Bryant Knob Fm c s
Noix Fm c s
Maquoketa Sh
Laketown Dol s c
?
Fish Haven Dol s c
Hanson c g s
?
Creek Fm c g s
Descon g
g
Fm

Fig. 5 Summary diagram showing important formations and degree of stratigraphical completeness of systemic boundary sections in nine important outcrop areas in the United States. Small letters, which indicate the presence of biostratigraphical control by means of a particular index fossil group, denote the following: c, conodonts; g, graptolites; s, shelly fossils (especially brachiopods); p, palynomorphs (especially chitinozoans). For further data on each of these successions, see the text. The section of northern Utah is that described by Leatham (1985). Vertical ruling marks proved or assumed stratigraphical gaps. Only formations near the systemic boundary are listed in the diagram.

cisely and broadly with the graptolite zone succession, and therefore these fossils cannot yet be used successfully to pinpoint the precise level of the Ordovician–Silurian boundary, especially in sections without a significant stratigraphical gap in the boundary interval. If the base of the *P. acuminatus* Zone is to be a viable and useful level for the base of the Silurian, then it is clearly necessary to determine the precisely equivalent level in the successions of shelly fossils, conodonts and palynomorphs. Because of the absence of graptolite control in the critical sections in the United States, that biostratigraphically most important correlation work will have to be carried out elsewhere in the world. However, the mutual stratigraphical relationships between non-graptolitic taxa are well displayed in sections in the United States. A detailed study of these relations no doubt will produce interesting and useful information of regional significance.

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The Ordovician–Silurian boundary in South America

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Synopsis

In South America late Ashgill rocks followed in the same succession by the early Llandovery are known only in the Precordillera of San Juan, Argentina. Early Llandovery fossils are known from the Puna Well, Argentina, the basal Trombetas Formation of Brazil, west of Lake Titicaca in Peru, and in the Merida Andes of Venezuela. Glaciogenic deposits of presumed Ordovician–Silurian boundary age are known from Argentina, Bolivia, Brazil and Peru.

Introduction

There are unfossiliferous and relatively unfossiliferous strata in South America whose assignment to either the Ordovician or Silurian is a problem. But I am unaware of any South American area where there are *fossiliferous* strata involved in real Ordovician–Silurian boundary indecision. In South America the assignment of fossiliferous beds to either the Ordovician or Silurian has been easy because there are no areas recognized to date where fossiliferous beds of latest Ordovician and earliest Silurian age are present in conjunction with each other.

Discussion of the Ordovician–Silurian boundary in South America may be broken into two parts: (1) the strata present on the shield areas; (2) the strata present in the structurally complex Andean regions bordering the shield areas to the north and west. Recognized, fossiliferous Ordovician rocks have not yet been shown to exist on the shield areas except for a few areas very close to the Andean, disturbed rocks, whereas there is widespread Ordovician scattered here and there in the Andean regions; fossiliferous Silurian rocks are widespread on the shield areas, as well as in the Andean regions. There are potentially Ordovician, unfossiliferous strata, possibly latest Ordovician, rocks on the shield areas, but until some means of dating them precisely emerges it would be futile to spend time discussing them. For example, Caputo & Crowell (1985) have described diamictites that may be tillites of Ashgill age, that occur not too far below Silurian strata containing higher land plant spores of earlier Llandovery age (Gray, unpublished data from the Amazon Basin). I will, as stated, not devote attention to such difficult and biostratigraphically ambiguous beds.

In the following summary statement I will review, geographic region by geographic region, what is currently known about the lowest Silurian and highest Ordovician fossiliferous rocks of the continent. It should, however, be kept in mind that the later Ordovician and earlier Silurian of South America are very poorly known, or known only in a rough reconnaissance manner, when contrasted with rocks of similar age in Europe. Conclusions arrived at here, particularly in the many poorly understood Andean regions, will certainly be subject to serious revision during the next few decades as additional field and laboratory studies take place.

The Silurian correlation chart for South America (Berry & Boucot 1972) provides a good summary of the data available up to about 1970, but can now be significantly supplemented by additional published and unpublished data. Extra new data are also published by Cuerda *et al.* and Baldis & Pöthe de Baldis (this volume, pp. 291–295).

Argentina

Amos (*in* Berry & Boucot 1972) provided an authorative review of the Argentinian Silurian, and its relations with the underlying Ordovician where present. The Argentinian Palaeozoic may be easily divided into that associated with the Andes in the north and the west, as contrasted with that present on the shield areas to the east. Much of the shield area Palaeozoic in Argentina is present in the subsurface beneath Mesozoic and Cenozoic cover, but there are

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limited areas where high-angle faulting has brought Precambrian and Palaeozoic rocks to the surface.

The shield regions in the Buenos Aires, La Pampa and Rio Negro regions (Amos *in* Berry & Boucot 1972: fig. 2) have not yielded any body fossils of proved Ordovician age, although some unfossiliferous units have been assigned for varied reasons to the Ordovician. Fossiliferous Silurian rocks are present in these regions, but no fossils of proved Lower Llandovery age have been demonstrated. The Silurian fauna consists of Malvinokaffric Realm brachiopods for the most part, and, as is characteristic of that cool to cold climate Realm, few taxa are present. It is presently unclear in these regions whether strata that could conceivably have crossed the Ordovician–Silurian boundary are present. The prevalence of late Ordovician to earlier Silurian continental glaciation in the Southern Hemisphere opens up the possibility that any such beds might well be in the non-marine category that can only be dated with a certain level of uncertainty for this time interval. The presence in the Cape Mountain System (Gray *et al.* 1986) of nearshore marine and possibly non-marine beds of probable Lower Llandovery or Ashgill age, or both, has some bearing on the Argentinian shield type occurrences in the Sierra de la Ventana, to the southwest of Buenos Aires in the Sierras Australes, which are commonly considered to be a pre-Jurassic continuation of the Cape Mountain System by many. In any event, it is reasonable to conclude (in the total absence of any dated Ordovician or early Llandovery fossils) that non-marine, or very nearshore, relatively unfossiliferous boundary beds might have been, or still might be present in the shield portions of Argentina. More subsurface data could demonstrate this possibility, particularly through the use of palynomorphs.

For purposes of considering the Ordovician–Silurian boundary, the Andean regions of Argentina should be divided into the Precordillera de San Juan, where the Cambrian and Ordovician fossils have North American platform biogeographical affinities and occur in platform carbonate type rocks, and the Andes proper with their Malvinokaffric Realm Ordovician and Silurian faunas occurring in siliciclastic rocks.

Amos (*in* Berry & Boucot 1972) has provided a summary for the Silurian of the Precordillera de San Juan. Nowhere are there fossiliferous Silurian rocks suspected to be older than Upper Llandovery, and the underlying Ordovician is nowhere thought to be younger than Caradoc, i.e. the Precordillera de San Juan is not a place in which to find a close approximation to the Ordovician–Silurian boundary as far as was then known, but see Cuerda *et al.* and Baldis & Pöthe de Baldis (this volume). The only exception to this statement about the absence of the Ashgill is in a limited area, where the Cantera Formation (Furque & Cuerda 1979: 473) has yielded Ashgill trilobites and brachiopods (Baldis & Blasco 1975; Nullo & Levy 1976; Levy & Nullo 1974), although interrupted above by '*contacto tectonico*' with a Lower Devonian unit. Tillites are not reported from this region, which suggests that the area may not necessarily have been subjected to continental glaciation, and might have been the site of a major regression associated with the terminal Ordovician–earliest Silurian glaciation.

Amos (*in* Berry & Boucot 1972) has summarized the Andean Silurian of northwestern Argentina, chiefly in the Provinces of Salta and Jujuy. The fossiliferous Silurian is no older than about Upper Llandovery based on available data, except for the single Lower Llandovery fossiliferous occurrence in the Puna well to the west of the material summarized by Amos (see Boucot *et al.* 1976). This fossiliferous Silurian is underlain by the tillites of the Mecoyita Formation which lack diagnostic fossils, and have been commonly considered (Laubacher *et al.* 1982) to be of Ashgill age (although shown by Amos, *in* Berry & Boucot 1972, to be well up into the Upper Llandovery). The underlying fossiliferous Ordovician is nowhere demonstrated to be of Ashgill age, although Caradoc equivalents are recognized (Amos *in* Berry & Boucot 1972).

It is clear that there are few places anywhere in Argentina for a palaeontologically-based close approach to the Ordovician–Silurian boundary.

Bolivia

Fossiliferous Ordovician (Hughes 1981, summary) and Silurian (Laubacher *et al.* 1982) rocks are well known in the Andean portions of Bolivia. However, no proved fossiliferous Silurian of

Lower Llandovery age is known, nor fossiliferous beds of Ashgill age. Tillite separating fossiliferous rocks belonging to the two systems is widespread. The oldest fossiliferous Silurian at present recognized is of Upper Llandovery age (Berry & Boucot 1972) from the Pojo region, where both brachiopods and graptolites provide the date. It is likely that there is a major, glacially correlated disconformity over most of Bolivia between the two systems (Berry & Boucot 1972: fig. 2). There is no reliable palaeontological evidence for placing any of the Andean tillites above the Llandovery: Berry & Boucot (1972: 26–27) summarize the graptolitic and brachiopod evidence from the overlying Kurusillas and Llallagua Formations, which contradicts that provided by Crowell *et al.* (1980); Crowell *et al.* (1981) suggest a Wenlock or Ludlow lower limit based on palynomorphs. An Ashgill age is most consistent for these tillites, in view of the overall emphasis on a glacial peak during that interval as contrasted with earlier Ordovician and later Silurian times. Antelo (1973) described Llandovery fossils from the Cancaniri, but the fossils actually come from above the tillite horizon (Cuerda & Antelo 1973) in beds which at Pojo were assigned by Berry & Boucot (1972) to the Llallagua Formation, which overlies the tillite proper.

Brazil

Fossiliferous Ordovician from the shield areas is unknown, except far to the west in the Amazonian region in the subsurface close to the areas of Andean disturbance. Silurian (Lange, *in* Berry & Boucot 1972) has been known from the Brazilian shield areas for over a century, but the graptolitic Silurian featuring *Climacograptus* has been conventionally assigned to the Llandovery, and not the latest Llandovery, because that genus was unknown above the Llandovery in the classic European and North American areas. Since 1972 there has been an accumulation of data indicating that *Climacograptus* can occur as high as the Lower Devonian (Jaeger 1978) in Austria, and that the palynomorphs associated with the graptolite show that the graptolites are no older than about Ludlow, rather than being of Llandovery age as had always been assumed. The palynomorphs in the Amazon Basin, where they occur with the graptolite, include acritarchs being studied by Luis Quadros, chitinozoans being studied by Florentin Paris, and higher land plant spores being studied by Jane Gray. All three specialists concur in assessing the age of the graptolitic part of the Trombetas Formation, the unit in question, as being no older than Ludlow. There is a possible tillite beneath the Trombetas Formation (Caputo & Crowell 1985). The tillite and associated strata are unfossiliferous, but an Ashgill age has been inferred, largely because the overlying, fossiliferous Trombetas Formation was concluded earlier to have been of Lower Llandovery age; this is now known to be an error. But basal Trombetas Formation beds, strata lacking any marine megafossils or marine palynomorphs, have yielded spore tetrads to Jane Gray which are of earlier Llandovery age and which also indicate in the absence of any marine organisms a possible non-marine environment. Similar spore tetrads of similar age have been recovered from the Brazilian Paraná Basin (Gray *et al.* 1985) and from the Cape Mountain System of South Africa (Gray *et al.* 1986).

Silurian strata have been reported from the Parnaiba Basin (Lange *in* Berry & Boucot 1972 gives a summary) based on palynomorph studies. However, there is still uncertainty about the precise parts of the Silurian present within this Basin, and no fossils of proved Ordovician age are known.

Fossiliferous Silurian was unknown in the Brazilian part of the vast Paraná Basin until this decade (see Gray *et al.* 1985, for a summary, including the initial recognition of these beds and their fossils by de Faria). Now, with the aid of both acritarchs and higher land plant spore tetrads there is no doubt about the presence of shallow water, Benthic Assemblage 1, marine earlier Llandovery on the northeastern flank of the Basin. Earlier Silurian, based on graptolites from the southwestern flank of the basin in Paraguay, has been known for some time (Harrington *in* Berry & Boucot 1972), but no trace of any fossiliferous Ordovician is known anywhere to be associated with the Paraná Basin.

In summary the Brazilian shield areas are not ones where the Ordovician–Silurian boundary may be located by means of fossils, owing to the total absence of any Ordovician fossils immediately beneath the available Lower Silurian fossils.

Chile

Fossiliferous Silurian rocks are unknown in Chile. The rocks from the Salar de Atacama region in northern Chile, assigned by Cecioni & Frutos (1975) to the Lower Palaeozoic (Ordovician, Silurian and Lower Devonian) are probably of Lower Carboniferous age, due to the similarity of their brachiopods to those found nearby (Bahlburg *et al.* 1986) which were assigned by Boucot to the Lower Carboniferous (fossiliferous Devonian beds are known from this area, yielding *Tropidoleptus* and *Australocoelia*, but these shells are unlike those figured by Cecioni & Frutos 1975 as contrasted with the Lower Carboniferous brachiopods). Fossiliferous earlier (Arenig) Ordovician is known in the Puna de Atacama, well to the east of the Salar de Atacama, but unassociated with fossiliferous Silurian. The nearest fossiliferous Silurian consists of a single Lower Llandovery locality in the Argentinian Puna, which yielded *Cryptothyrella* among other things (Boucot *et al.* 1976), which is unassociated with any fossiliferous Ordovician. The fossiliferous Devonian beds in the Salar de Atacama region are no older than about Siegenian–Emsian, and rest unconformably on an older basement complex. We do not know whether there is any possibility of finding Ordovician–Silurian boundary region strata in Chile. The older Palaeozoic rocks of Chile are almost unknown, although there are many suspect regions that warrant careful attention.

Colombia

Fossiliferous Silurian rocks are presently unrecognized in Colombia, while none of the known Ordovician has been shown to even reach the Caradoc, much less the Ashgill (Hughes 1981). The presence in the Perija Andes, on the Colombian–Venezuelan boundary, of Lower Devonian fossiliferous beds, resting unconformably on basement complex, indicates that at least in some spots one would not expect fossiliferous Silurian or Ordovician strata to be preserved.

Ecuador

Fossiliferous Ordovician and Silurian rocks have not yet been recognized in Ecuador, although there is no reason to doubt their potential presence in the Andean part of the country.

Paraguay

See discussion of the Paraguayan Lower Silurian occurring on the southwestern margins of the Paraná Basin under 'Brazil', p. 287.

Peru

There is widespread fossiliferous Ordovician and Silurian in southern Peru, both to the east and west of Lake Titicaca (see discussion of the Silurian in Laubacher *et al.* 1982; Hughes, 1981, summarizes the Ordovician, which has reliable palaeontological evidence only up to beds of Caradoc age). Laubacher *et al.* (1982) recognized Early Llandovery brachiopods to the west of Lake Titicaca, in the absence of the tillite that so commonly separates fossiliferous Ordovician and Silurian rocks from each other in the central Andean region. But these fossiliferous Early Llandovery fossils are removed stratigraphically some distance from the youngest Ordovician rocks which have yielded fossils no younger than Caradoc. In southern Peru, therefore, there is no locality known where a close approach to the Ordovician–Silurian boundary is made within fossiliferous rocks. In central and northern Peru, as well as along the coast, fossiliferous Silurian rocks are unrecognized. The lack of tillite to the west of the Titicaca region does raise the possibility that an Ordovician–Silurian transition may eventually be discovered in southern Peru or adjacent Bolivia, since a major disconformity might be more likely in the more easterly regions characterized by tillite.

Venezuela

The Ordovician and Silurian rocks related to the Ordovician–Silurian boundary are restricted in their occurrence to the Merida Andes, well to the south of Lake Maracaibo. Hughes (1981) comments that the Ordovician faunas of the Merida Andes are of Caradoc age; they are structurally well removed by faulting from immediate contact with the Lower Llandovery faunas of the Merida Andes described by Boucot *et al.* (1972). The Lower Llandovery faunas of the Merida Andes are dominated by brachiopods that cannot be dated any closer than Lower Llandovery; thus we are ignorant about whether or not these faunas are actually very close to the Ordovician–Silurian boundary. Graptolites that might help to resolve the age problem are unknown from the Merida Andes Llandovery. The shallow water nature of the Merida Andes Lower Llandovery, the medium-grained sandstones of the Silurian portion of the Caparo Formation with a Benthic Assemblage 2 set of communities dominated by such genera as *Mendacella*, is, however, consistent with the concept that there might be a disconformity between the two systems there, related to possible glacial regression, as is the case in many other parts of the world. In any event, the recognition of a close approximation to the Ordovician–Silurian boundary in Venezuela is as yet unknown.

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The Ordovician–Silurian boundary in Bolivia and Argentina

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Synopsis

The Ordovician–Silurian boundary level has been identified in few areas, although there is considerable potential for future work. The following sections are the best: 1 Lampaya, Bolivia; 2 the Don Braulio Valley, Argentina; 3 Talacasto, Argentina. Recent fieldwork has established that Talacasto appears the best of these, and a sequence of *persculptus* Zone, probably *acuminatus* Zone, and approximate equivalent of the *atavus* Zone has been established. The base of the Silurian at Talacasto is taken at 60 cm above the base of the La Chilca Formation, following a *persculptus* Zone assemblage. Several stratigraphically important graptolites are recorded from South America for the first time.

Introduction

In Bolivia undoubted low Silurian rocks are exposed in the Eastern Cordillera, and in Argentina in the Precordillera (Fig. 1). The Cancañiri Formation is the basal unit of the Silurian in Bolivia (Castaños & Rodrigo 1978) and consists of 105 m of diamictites, shales and sandstones yielding palynomorphs and, in some sections, scarce brachiopods. The Precordilleran Argentinian Silurian is recognized as three facies types: the Eastern Facies, some 2500–3000 m of shales, sandstones and conglomerates with associations of brachiopods, corals and graptolites; the Central Facies, 450–500 m of green shales, orthoquartzites, and fine grained limestones, with rich assemblages of brachiopods, corals, trilobites and graptolites; and the Western Facies, restricted to the Calingasta region, approximately 1000 m of shales and turbidite sandstones, yielding some brachiopods. Each facies type (Cuerda, in press) is interpreted as having a different palaeoenvironment, respectively: a N–S trough between Pre-Cambrian ridges; proximal to distal platform; distal platform to abyssal plain. The stratigraphically lowest formations in these facies are the La Rinconade Formation, the La Chilca Formation, and the Calingasta Formation.

Bolivia

The Lampaya section is located near Cochabamba. Three lithological units have been recognized in the Silurian, the Cancañiri Formation at the base, and above it the Kirusillas and Catavi Formations, a total of 1355 m spanning the Llandovery to Ludlow. The Ashgill Series seems to be absent in Bolivia so that the Cancañiri Formation rests upon Caradoc or earlier strata. At Lampaya the Cancañiri Formation consists of 105 m of diamictites with shales and sandstones intercalated as thin layers. A Llandovery age is supported by palynomorphs referable to the *Veryhachium rhomboidium* Zone (Suarez-Riglos 1975). Macrofossils have been recovered including trilobites, brachiopods, corals and ostracods by one of us (A.C.). The Cancañiri Formation at Lampaya rests upon the Caradoc.

Argentina

Villicum Hills Section. The Don Braulio Valley drains the eastern slopes of the Villicum Hills, where the Ashgill black shales and grey sandstones are topped by a ferruginous oolite. The grey

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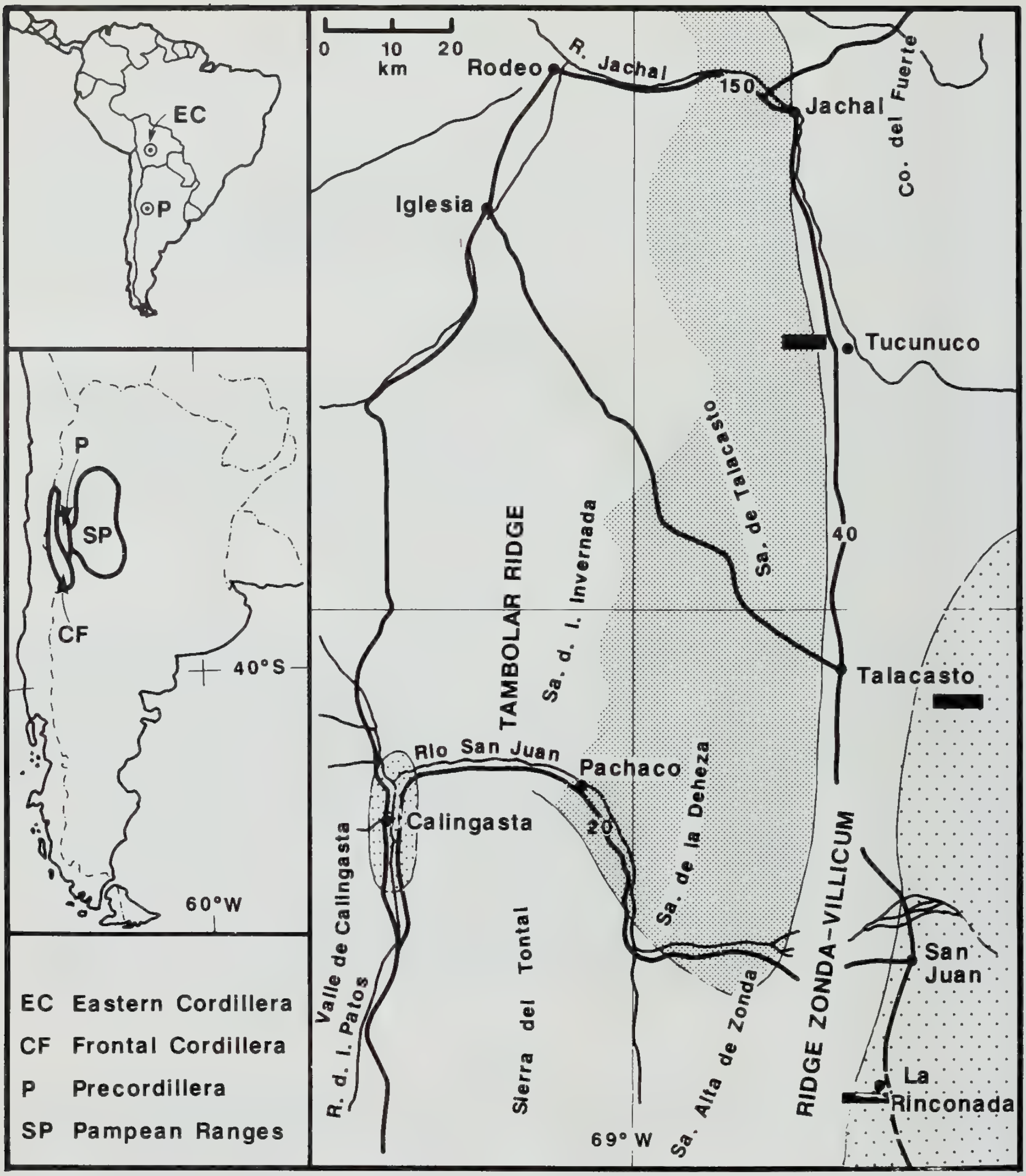
EC
P
SP
CF
40°S
60°W
EC Eastern Cordillera
CF Frontal Cordillera
P Precordillera
SP Pampean Ranges
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R. Jachal
150
Jachal
Co. del Fuerte
Iglesia
Tucunuco
Sa. de Talacasto
40
TAMBOLAR RIDGE
Sa. d. l. Invernada
Talacasto
Rio San Juan
Pachaco
Calingasta
20
Sa. de la Deheza
RIDGE ZONDA–VILLICUM
Valle de Calingasta
R. d. l. Patos
Sierra del Tontal
Sa. Alta de Zonda
San Juan
La Rinconada
69° W

Fig. 1 Distribution of Silurian facies in the Precordillera of San Juan, Argentina. The western facies is shown around Calingasta, the central facies in the close stipple and the eastern facies in open stipple to the right.

sandstones have yielded the trilobites *Calymenella* (*Eohomalonotus*) *villicumensis* Baldis & Blasco and *Dalmanitina sudamericana* Baldis & Blasco (Baldis & Blasco 1974) and the brachiopods *Fascifera punctata*, *Arenorthis cuyana*, *Villiscundella muozetici*, *Bagnorthis garrigoui* and *Kjaerina* (*Neokjaerina*) *florentina* (all Levy & Nullo 1977).

The Silurian commences with argillaceous sandstones and has a palynomorph assemblage referable to the Llandovery, which Volkheimer *et al.* (1980) list as *Ancyrochitina* sp., *A.* cf.

ancyrea (Eisenack), *Conochitina* cf. *chydaea* Jenkins, *Desmochitina* sp., *Cyathochitina* cf. *campanulaeformis* Eisenack, *Euconochitina* cf. *filifera* Tangourdeau, *Rhabdochitina* sp. 'A', *Spathochitina* cf. *clarindoi* de Costa and *Sphaerochitina* sp. Above the argillaceous sandstones the beds grade into medium and coarse sandstones of Wenlock and Ludlow age (Magotes Negros Formation). Baldis & Pöthe de Baldis (1988, this volume) have reviewed and revised this section.

The **Talacasto** section (Figs 1, 2) is located some 16 km WNW of Talacasto railway station and has been studied by Cuerda *et al.* (1982). Recent collecting by the authors yielded several hundred graptolites throughout the whole of the 3·65 m of the La Chilca Shale Formation. Collecting was done every few centimetres, as closely as the friability of the shale would allow. Several confirmatory collections were made nearby. *Glyptograptus persculptus* occurs commonly, both flattened and in three dimensions, in association with equally common specimens of *Climacograptus angustus* Perner and in addition *Pseudoclimacograptus* sp. nov., *Glyptograptus* sp. (an undescribed form commonly seen in the *persculptus* Zone in other parts of the world), *Climacograptus* cf. *medius* Törnquist, and *Climacograptus normalis* Lapworth. This assemblage is taken to indicate the latest Ordovician *G. persculptus* Zone.

At 55 cm above the base of the formation *G. persculptus* s.s. disappears, but the remainder of the fauna continues. *Rhaphidograptus* sp. at 90 cm, and *G.* ex gr. *persculptus* (late forms, smaller, and with a delayed median septum) also occur between 1·1 m and 1·38 m, where *Pseudoclimacograptus* sp. nov. is also especially abundant and dominates the fauna. The *Pseudoclimacograptus* sp. nov. is close to *P. fidus* and *P. pictus* described from the *acuminatus* Zone of Kazakhstan by Koren & Mikhailova (1980). From 60 cm to 1·7 m we have recorded specimens

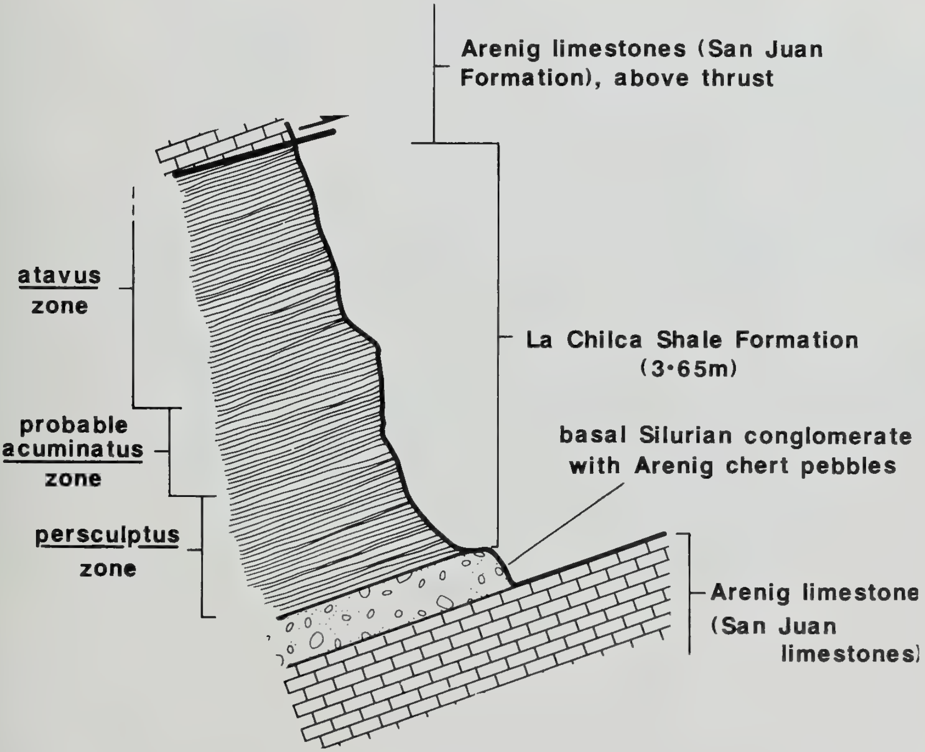
Arenig limestones (San Juan Formation), above thrust
atavus zone
probable acuminatus zone
persculptus zone
La Chilca Shale Formation (3·65m)
basal Silurian conglomerate with Arenig chert pebbles
Arenig limestone (San Juan limestones)

Fig. 2 Section through the Ordovician–Silurian boundary near Talacasto, San Juan Province, Argentina. The 'basal Silurian conglomerate' also includes the *persculptus* Zone.

of *Climacograptus acceptus* Koren & Mikhailova, also typical of the *acuminatus* Zone, and we have found specimens possibly referable to *Glyptograptus maderni* Koren & Mikhailova from 60–90 cm. At 1·6 m there is a further change in the fauna, with the disappearance of glyptograptids and the *Pseudoclimacograptus*, whilst there is an increase in abundance of *C. angustus*, *C. normalis* and *C. rectangularis* and the appearance for the first time of the monograptid *Lagarograptus. Paraclimacograptus* cf. *innotatus* (Nicholson) appears at 1·75 m. This fauna is then maintained to the top of the section apart from the addition of a new diplograptid.

The base of the *acuminatus* Zone, and hence of the Silurian, is probably best taken at 60 cm with the appearance of *Climacograptus acceptus*. For reasons which we shall discuss in a systematic paper elsewhere, we take the incoming of *Lagarograptus* to be roughly equivalent to the *atavus* Zone.

Thus the Talacasto region at present affords the best recognition of the base of the Silurian in South America. The potential is considerable for further precise subdivisions on other sections in the same region. The authors' recent fieldwork established the following stratigraphically important forms for the first time in South America: *G. persculptus*, *C. angustus*, *C. normalis*, *C. acceptus*, *C. rectangularis*, *Rhaphidograptus*, *Paraclimacograptus*, and *Lagarograptus*.

Acknowledgements

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The Ordovician–Silurian boundary in the Sierra de Villicum, Argentine Precordillera

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Synopsis

The Ordovician–Silurian boundary is defined within the Don Braulio Formation at its type locality near San Juan, Argentina. The boundary sequence consists of: 1, Upper Ashgill (Hirnantian) defined by the presence of *Hirnantia* cf. *sagittifera* and *Dalmanitina sudamericana*; 2, a short stratigraphical interval of 10 m of shales with unidentifiable graptolite fragments, perhaps Lower Silurian in age; 3, levels with acritarchs, chitinozoan and graptolites which can be related with certainty to the Lower Llandovery.

The stratigraphical section which includes the Ordovician–Silurian boundary in the Sierra de Villicum is perhaps the best known and palaeontologically controlled locality in South America. The Sierra (Range) of Villicum is situated in the Argentine Precordillera, in San Juan Province about 1100 km northwest of Buenos Aires (see Fig. 1). Upper Ordovician and Silurian sediments outcrop in the eastern flank of the range, and the best section is found in Don Braulio Creek, 35 km north of the city of San Juan. The section is well exposed, in a desert climate area, and the following formations are present:

Mogotes Negros Fm	Lower to Upper Silurian age
Don Braulio Fm	Ashgill to Llandovery age
La Cantera Fm	Llandeilo to Caradoc age
Los Azules Fm	Llanvirn to Llandeilo age
San Juan Fm	Arenig age
La Flecha Fm	Upper Cambrian to Tremadoc age
La Laja Fm	Lower to Middle Cambrian age

The first Ashgill macrofossils from South America were found in the Don Braulio Formation (Baldis *et al.* 1982). The brachiopods were described by Levy & Nullo (1974) and trilobites of the *Dalmanitina* faunal group by Baldis & Blasco (1975). Benedetto (1985) has reported the presence of *Hirnantia* associated with *Modiolopsis* (Sanchez 1985), which gives an accurate Upper Ashgill age for the top of the lower Don Braulio Formation.

The trilobites found in the lower part of the formation are *Dalmanitina* (*D.*) *sudamericana* Baldis & Blasco and *Calymenella* (*Eohomalonotus*) *villicunensis* Baldis & Blasco, and brachiopods belonging to the genera *Fascifera*, *Arenorthis*, *Bagnorthis* and *Kjaerina* (*Neokjaerina*). From the middle to the upper part of the lower portion of the formation are reported *Hirnantia sagittifera* (M'Coy) and *Dalmanella* aff. *D. testudinaria*, associated with *Modiolopsis*, *Nuculopsis* and *Palaeoneilo*. The lower part of the formation is separated by several metres of shales with indeterminable remains of graptolites from the upper part.

In the base of the upper part of the formation, Volkheimer *et al.* (1980) determined a chitinozoan microflora composed of *Ancyrochitina* cf. *ancyrea* (Eisenack) Eisenack, *Conochitina* cf. *chydae* Jenkins, *Desmochitina* (?) sp., *Cythochitina* cf. *campanulaeformis* (Eisenack) Eisenack, *Euconochitina filifera* (Eisenack) Tang, *Rhabdochitina* sp. A, and *Spathachitina* cf. *clarindoi* da Costa. The Llandovery age of the association is indicated by the presence of Ordovician–Lower Silurian chitinozoa together with Lower Silurian ones. The genus *Spathachitina* da Costa indicates a Lower Silurian age in the Amazon Basin of Brazil. Pöthe de Baldis (1980) described a varied microflora of acritarchs from the same level, with 26 genera and 47 species, of which 34 are known from other countries, mainly northern Spain, Belgium, England and northern Africa. The association shows a predominance of *Veryhachium trispinosum*, followed in importance by *Eupoikilofusa tenuistriata* (Pöthe de Baldis) *aperturata* n. var. The genus *Eisenackidium*

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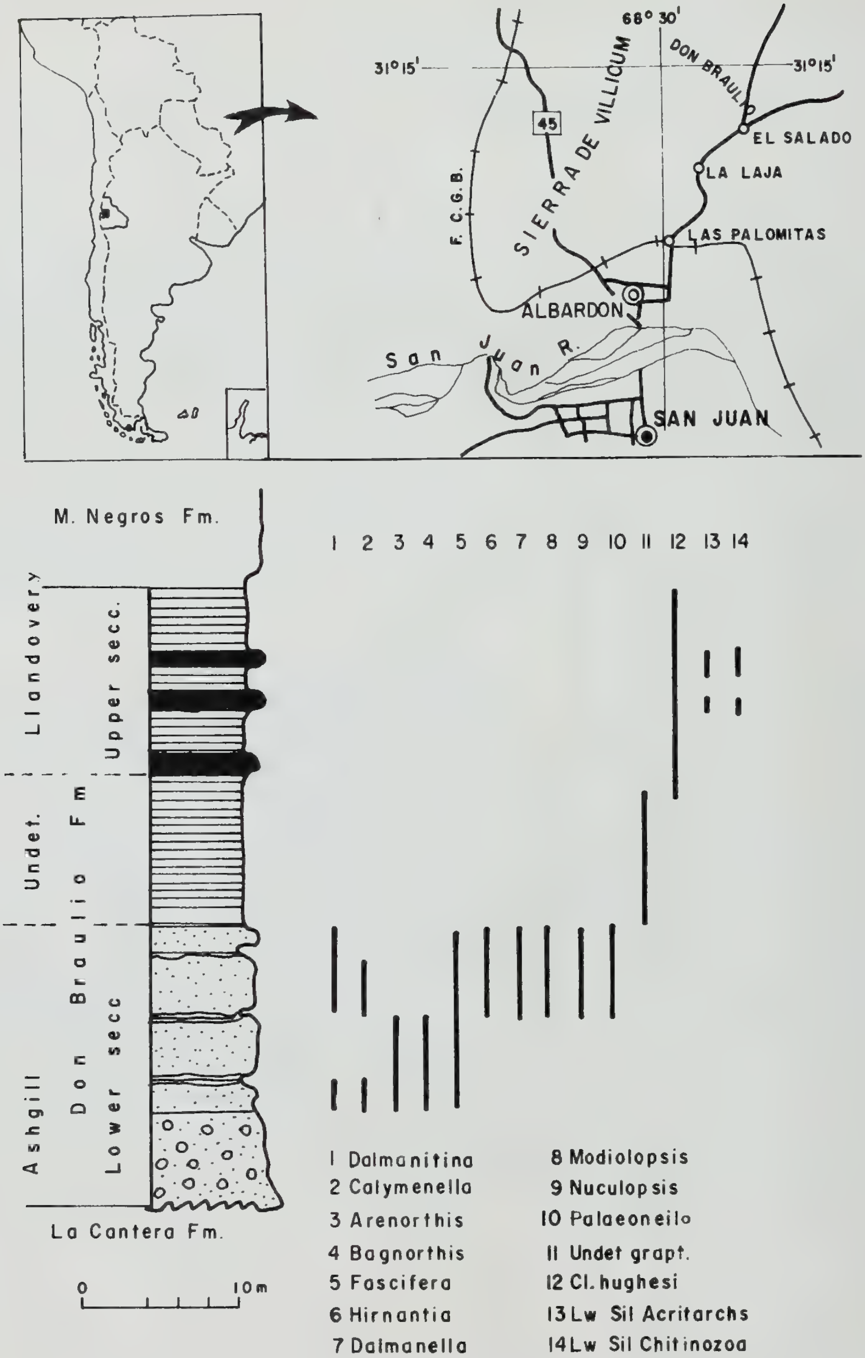
68° 30'
31°15'
31°15'
DON BRAULIO
SIERRA DE VILLICUM
45
F. C.G.B.
EL SALADO
LA LAJA
LAS PALOMITAS
ALBARDON
San Juan R.
SAN JUAN
M. Negros Fm.
Llandovery
Upper secc.
Undet.
Don Braulio Fm
Ashgill
Lower secc
La Cantera Fm.
0
10m
1 2 3 4 5 6 7 8 9 10 11 12 13 14
1 Dalmanitina
2 Calymenella
3 Arenorthis
4 Bagnorthis
5 Fascifera
6 Hirnantia
7 Dalmanella
8 Modiolopsis
9 Nuculopsis
10 Palaeoneilo
11 Undet grapt.
12 Cl. hughesi
13 Lw Sil Acritarchs
14 Lw Sil Chitinozoa

Fig. 1 The location of the type locality of the Don Braulio Formation (above), and a section through the formation showing the distribution of the fossils mentioned in the text (below).

was recorded for the first time from the Lower Silurian (formerly only described from the Lower Devonian). Other forms such as *Veryhachium tetraedron* Deunff, *Marrocanium simplex* Cramer *et al.*, *Multisphaeridium alloiteaui* Deunff and *M.* cf. *remotum* (Deunff) are typical of Ordovician sediments. The age of the association is based on the presence of *Tunisphaeridium tentaculaferum* (Martin) and *Domasia limaciforme* (Stockman & Willière) whose first appearance is in the Lower Llandovery of England and Belgium.

The graptolites appearing in all parts of the Upper Section of the Don Braulio Formation were determined by Peralta (in press) as the typical Lower Silurian assemblage of *Climacograptus* aff. *C. hughesi* (Nicholson), *Monograptus* sp., *Glyptograptus* sp. and *Rastrites* sp.

The Don Braulio Formation has 40 m thickness in the type locality. A brief description of the section is as follows (also see Fig. 1):

A.	Hematitic Member:	
9	Pale green-greyish shales (strongly deformed), with spots of iron oxides and scattered ramous graptolites	2 m
8	Oolitic sandstones weathering dark red; red brownish to green in fresh fracture. Acritarchs and chitinozoa occur	0·50 m
7	Pale green-greyish shales, intercalated with thin hard siltstones and fine-grained sandstones of 2 cm thickness	2 m
6	Oolitic sandstone similar to Bed 8 but with fewer oolites, with indeterminable monograptid stipes	1 m
5	Green-greyish hard siltstone with some shaley levels and *Climacograptus hughesi*	3 m
4	Hematitic siltstones (poorly bedded)	1 m
B.	Silty and Shaly Member:	
3	Dark green shales and siltstones (highly deformed) alternating with pale green-greyish clay, with fragments of unidentifiable graptolites (Monograptidae?)	10 m
C.	Conglomerate and Sandstone Member (Lower Section):	
2	Dark green to green-greyish fine-grained sandstones (poorly bedded) with the trilobites *Calymenella* (*Eohomalonatus*) *villicumensis*, *Dalmanitina sudamericana* and the brachiopods *Fascifera punctata*, *Hirnantia* cf. *sagittifera*, *Dalmanella* cf. *testudinaria*	12 m
1	Basal oligomictic conglomerate with intercalations of green-greyish lenses	6 m

Unconformity ~ ~ ~ ~ ~ ~
La Cantera Formation (Llanvirn to Caradoc).

From the above we may conclude that the Ashgill is well dated in the whole sequence with trilobites, brachiopods and *Hirnantia* cf. *sagittifera* at its top. Ten metres of barren shales follows this section, followed by a Llandovery graptolite fauna and acritarchs and chitinozoans of Lower to Middle Llandovery age.

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Late Ordovician and Early Silurian Acritarchs

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Synopsis

The principal stratigraphical data for late Ordovician and early Silurian acritarchs are reviewed; at present they do not justify any formal zonation on a broad geographic scale. The systemic basal boundary stratotype at Dob's Linn, southern Scotland, has not yielded index acritarchs. A preliminary selection of taxa from correlative strata on Anticosti Island, Québec, eastern Canada, indicates that the area has the most continuous palynological record from at least the Ashgill to the late Llandovery, with the best potential for establishing detailed acritarch systematics and interregional correlation.

Introduction

In general, the biostratigraphical tool provided by the acritarchs is still only partly exploited for interregional correlation, for the following reasons: (i) sufficiently detailed systematic descriptions have become available only during the last fifteen years or so, through the use of SEM, and a coherent taxonomic framework is still lacking; (ii) precisely defined taxa are most often reported only from regions where their total stratigraphical range is not established; (iii) a large number of data relate to dispersed samples, for which there is no macrofossil age control. In particular, acritarchs of latest Ordovician and earliest Silurian age have received little documentation. This scarcity of data reflects the lack of palynological investigations rather than of suitable marine deposits, for these probably planktonic, organic-walled microfossils appear to be relatively weakly facies-controlled when compared with macrofossils. Nevertheless, the Ashgill extinction that affected numerous other fossil groups also involved the acritarchs. Differences in composition of assemblages between the end of the Ordovician and the beginning of the Silurian are indicated in the following areas: Anticosti Island, eastern Canada; southern Appalachians, U.S.A.; Belgium; and the Algerian Sahara. These differences are amplified by the absence of Hirnantian or Gamachian strata, except on Anticosti, where, on the basis of preliminary data (Duffield & Legault 1981, and author's personal observations), the disappearance of numerous Ordovician taxa seems to occur in the Gamachian. A marked change between acritarch associations from the late Ashgill and the Llandovery is mentioned briefly (Le Hérissé 1984) for the subsurface rocks in southern Gotland. Colbath (1986) has reviewed different hypothetical causes for these acritarch extinctions, ranging from the effects of sea-level and climatic changes associated with glaciation to a bolide impact model.

Review of data

The map (Fig. 1) shows the distribution of late Ordovician and early Silurian acritarchs and indicates detailed references. Numbers (see explanation of Fig. 1) refer generally to the most recent publication that indicates previous data; exceptions are Anticosti and Great Britain, for which further references are given. Anticosti and southern Scotland provided the two final candidate sections for the Ordovician–Silurian boundary stratotype considered by the Subcommission on Silurian Stratigraphy (Holland 1984). Since then the International Commission on Stratigraphy (Bassett 1985) has chosen to fix the base of the Llandovery Series, together with that of its lowest stage, the Rhuddanian, at Dob's Linn, southern Scotland; the boundary stratotypes for the two other Llandovery stages, Aeronian and Telychian, are located in the type area of the Llandovery in Wales (Cocks 1985).

Areas from which no index acritarchs are known (for example, the Ashgill of southwest France, Rauscher 1974) are omitted. Owing to the lack of agreement on precise correlation between the North American and British upper Ordovician standard successions (Barnes *et al.*

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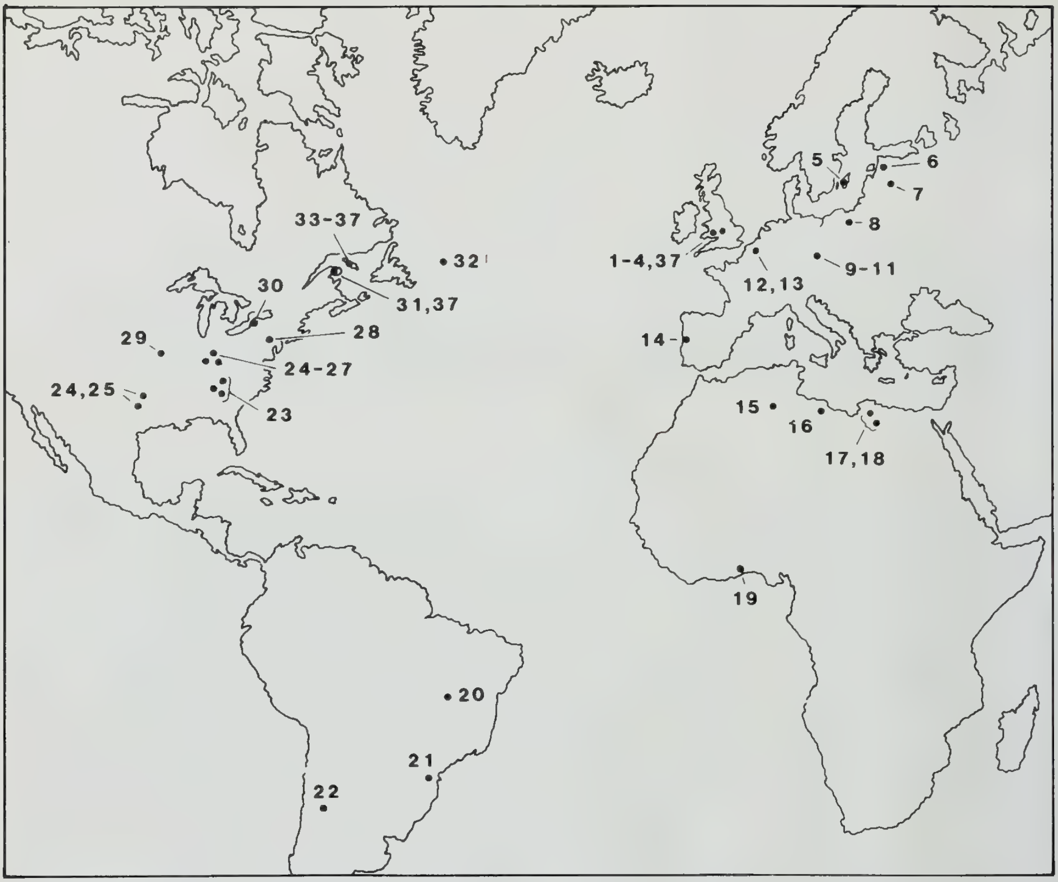


Fig. 1 Generalized world map showing late Ordovician and early Silurian acritarch localities. The following abbreviations indicate the information included in publications 1–37 listed below: (CA), undifferentiated late Caradoc and Ashgill; A, Ashgill; P, Pusgillian; R, Rawtheyan; H, Hirnantian; G, Gamachian; L, undifferentiated Llandovery, possibly including Rhuddanian; Rh, Rhuddanian; Ae, Aeronian; T, Telychian; p.d., palynological dating only. Chronostratigraphic units groups within parentheses are not differentiated from each other.

1, Hill 1974, Rh-T: 2, Aldridge *et al.* 1979, Rh-T: 3, Hill & Dorning *in* Cocks *et al.* 1984, Rh-T: 4, Downie 1984, Rh-T: 5, Eisenack 1968, A: 6, Eisenack 1963, A: 7, Umnova 1975, A, L: 8, Górka 1969, A: 9, Konzalová-Mazanková 1969, (PR): 10, Vavrdová 1974, A: 11, Vavrdová 1982, H: 12, Martin 1969, (CA), Rh-T: 13, Martin 1974, (CA), Rh: 14, Elaouad-Debbaj 1981, (CA), A, p.d.: 15, Jardiné *et al.* 1974, (C ?A), (AeT), p.d. in part: 16, Deunff & Massa 1975, ?C, p.d., ?Rh: 17, Molyneux & Paris 1985, A, p.d.: 18, Hill *et al.* 1985, (RhAe), p.d.: 19, Bär & Riegel 1980, (AL), p.d.: 20, Brito 1967, L, p.d.: 21, Gray *et al.* 1985, L, p.d.: 22, Melendi & Volkheimer 1985, L: 23, Colbath in press, (CA), (PR), (RhAe): 24, Loeblich & Tappan 1978, (CA), (PR): 25, Loeblich & McAdam 1971, (CA), (PR): 26, Loeblich 1970, (PR): 27, Colbath 1979, (CA): 28, Johnson 1985, L: 29, Wright & Meyers 1981, (CA), p.d.: 30, Miller & Eames 1982, Rh: 31, Martin 1980, (PR): 32, Legault 1982, (CA), p.d.: 33, Staplin *et al.* 1965, (PR): 34, Cramer 1970, (AeT): 35, Duffield & Legault 1981, 1982, G, Rh-T: 36, Jacobson & Achab 1985, (PR): 37, Martin in press and personal observation, G (at Anticosti only), Rh-T. [Since submission of this paper, Whelan (1986) has commented briefly on the acritarchs from Dob's Linn.]

1981; Ross *et al.* 1982; Shaver 1985), palynological references for both late Edenian and Maysvillian strata in U.S.A. are included. In the Llandovery Series, acritarch data given for the Rhuddanian sometimes include those for the Aeronian and Telychian. Localities where the sections begin only with the Aeronian or Telychian are omitted here and may be found in Martin (in press).

Europe

In Great Britain, no palynological work has been published on the Ashgill. The Ordovician–Silurian boundary stratotype strata at Dob's Linn (Cocks 1985) are composed of condensed, deep-water, graptolitic shales, the base of the Llandovery being coincident with the base of the *P. acuminatus* Zone. The whole succession, from the *Climacograptus peltifer* Zone (early Caradoc) upwards, contains rare, blackish acritarchs, but these are too poorly preserved to provide useful information. The type Hirnantian (Hirnant Limestone) at Cwm Hirnant quarry, near Bala, North Wales, yielded rare acritarchs belonging to either poorly-defined or remnant Arenig–Llanvirn taxa (personal observation). The Caradoc Series (Costonian to Onnian stages) in the type area of Shropshire contains well preserved assemblages (Turner 1984) of Caradoc age, associated with others derived from Tremadoc and Arenig–Llanvirn deposits. Rhuddanian microfloras from near Llandovery are both poorly preserved and of low diversity but permit (Hill & Dorning in Cocks *et al.* 1984) the recognition of three biozones characterized, on the basis of published lists, by the successive appearance of taxa that, for the most part, are long-ranging in the Silurian or are left in open nomenclature. The top of the Rhuddanian there also contains reworked, pre-Caradoc Ordovician material (Martin in press).

In the same region, and especially in the the Welsh Borderland (Hill 1974), partly published results for the Llandovery show, from the Aeronian onwards, a refined palynological zonation that may be compared with that outlined for Belgium (Martin 1969). Of particular significance are species of *Domasia* Downie, 1960 emend. Hill, 1974 and *Dilatisphaera williereae* (Martin) Lister, 1970.

In the Massif of Brabant, Belgium (Martin 1974), moderately well preserved acritarchs, mostly long ranging and including some known from the Tremadoc to the Arenig–Llanvirn, are from boreholes. Parts of these rock successions are assigned a late Caradoc and/or Ashgill age on lithological and structural grounds in the absence of diagnostic macrofossils; those of the basal Rhuddanian are dated by graptolites and include strata of the *P. acuminatus* Zone.

In the Baltic region (Gotland, Estonia, Latvia—Eisenack 1963, 1968; Umnova 1975), Poland (Górka 1969) and Czechoslovakia (Konzalová-Mazanková 1969; Vavrdová 1974, 1982), as in Portugal (Elaouad-Debbaj 1981), data are relatively few for the Ashgill and absent for the Rhuddanian. The only Hirnantian acritarchs so far illustrated come from the Prague region (Vavrdová 1982).

Africa and South America

Microfloras from boreholes in north Africa are well preserved. At the Grand Erg Occidental in the Algerian Sahara (Jardiné *et al.* 1974) acritarch zone F corresponds to the Caradoc and perhaps Ashgill; it also contains taxa characteristic of the Arenig–Llanvirn and is present too in deposits of the Illizi Basin attributed doubtfully to the *M. sedgwickii* Zone of the Aeronian. In Libya (Deunff & Massa 1975; Molyneux & Paris 1985; Hill *et al.* 1985) acritarchs from the late Ordovician and early Silurian, cited and partially figured, are dated with particular reference to palynological data from western Europe and central U.S.A. In Deunff & Massa (1975) the list of taxa alleged to have been found in the early Rhuddanian *C. vesiculosus* Zone indicates a post-Llandovery age and is not considered further here.

Acritarch data for the relevant interval in Ghana (Bär & Riegel 1980), Brazil (Brito 1967; Gray *et al.* 1985) and Argentina (Melendi & Volkheimer 1985) are dispersed and mainly without independent age control. The most noteworthy illustrated observation is that samples from Ghana said to occur at the Ordovician/Silurian boundary share only a single species, *Dactylofusa marahensis* Brito & Santos, 1965, with strata of the Maranhão Basin attributed to the Lower Silurian. In both cases the age is based on structural and palynological arguments.

North America

Publications referring to the eastern and central U.S.A. deal mainly with numerous new late Ordovician taxa from Oklahoma (Loeblich & McAdam 1971; Loeblich & Tappan 1978) and the Cincinnati area (Loeblich 1970; Loeblich & McAdam 1971; Loeblich & Tappan 1978; Colbath 1979); however, the acritarchs from the Richmondian Stage, which is correlated with part of the Ashgill Series, are from isolated samples. In the southern Appalachians (southwest

Virginia, northwest Georgia and east Tennessee), a consistent acritarch correlation, based largely on new taxa, is documented (Colbath, in press) for the passage from Ordovician to Silurian; but the presence of the Gamachian and earliest Rhuddanian in the region is debatable. An acritarch assemblage of undoubted Rhuddanian age in western New York State (Miller & Eames 1982) enables preliminary correlations to be made with assemblages in the southern Appalachians, Anglo-Welsh area and Belgium. A very few Llandovery, including perhaps Rhuddanian, acritarchs are known from central Pennsylvania (Johnson 1985).

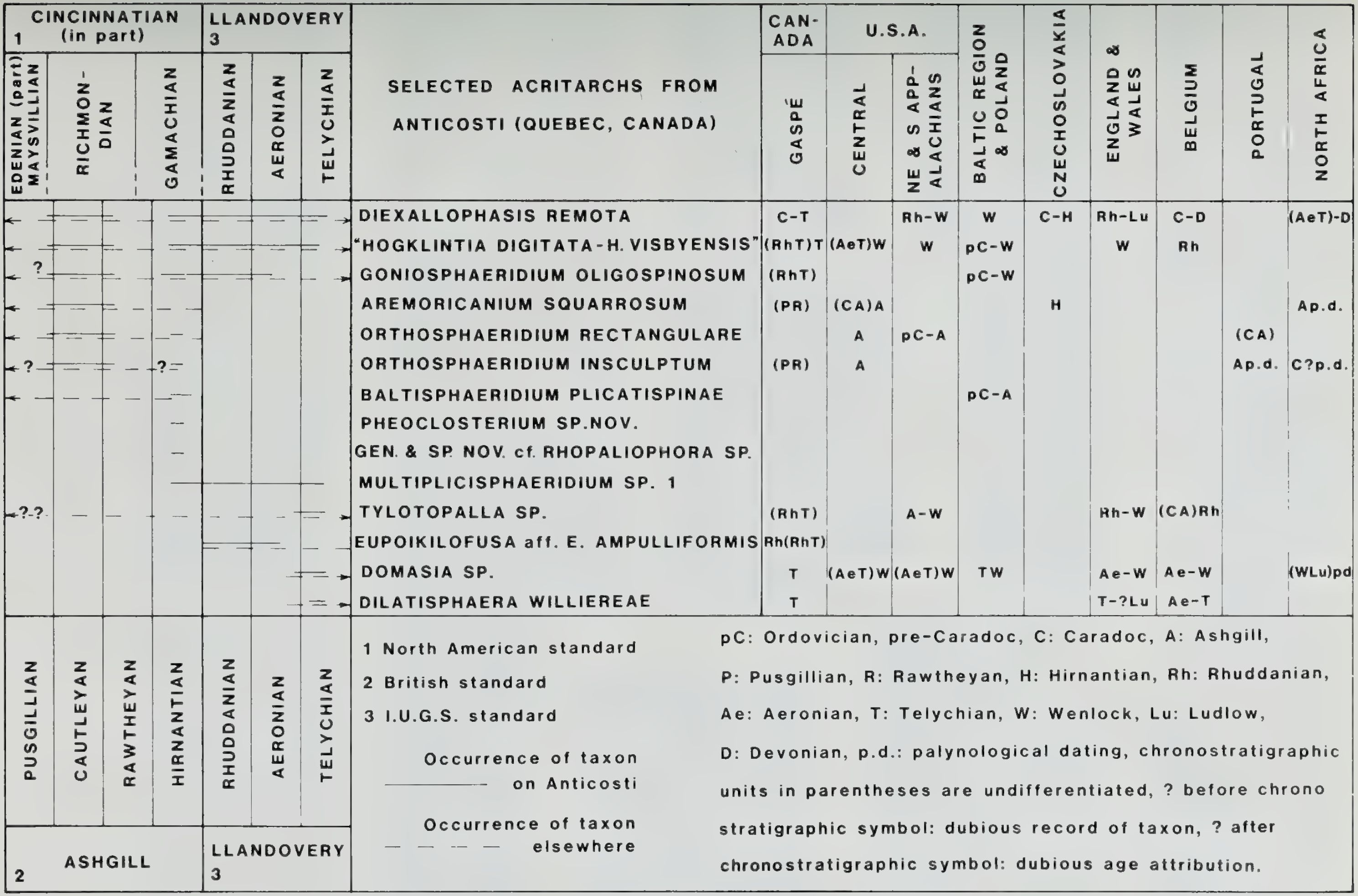
In eastern Canada, except for palynologically dated latest Caradoc or Ashgill strata in a borehole in the Labrador Sea (Legault 1982), data relate to the Province of Québec. Only reconnaissance studies are available for the pre-Hirnantian Ashgill of the Percé area (Martin 1980) in the Gaspé Peninsula. The White Head Formation at White Head (Lespérance 1985; Fig. 2 herein) has not yielded index acritarchs in the Hirnantian interval, and the basal Llandovery portion (base of Unit 6; personal observation) contains specimens deformed by crystal growth; some of the latter, for example *Eupoikilofusa* aff. *E. ampulliformis*, *sensu* Duffield & Legault 1981, are very characteristic of the Rhuddanian at Anticosti, from the base upwards of the Becscie Formation at Ellis Bay.

At Anticosti an Ordovician/Silurian boundary stratotype was proposed (Barnes & McCracken 1981) in an allegedly continuous limestone-shale succession in the upper part of the Ellis Bay Formation (*sensu* Petryk 1981) at Ellis Bay. The base of the Silurian is marked by the appearance of the conodont *Ozarkodina oldhamensis* (Rexroad, 1967); *Oulodus*? *nathani* McCracken & Barnes, 1981 is an auxiliary indicator for the boundary. However, Lespérance (1985) places the boundary higher and in the Becscie Formation, on the assumption that the appearance of the trilobite *Acernaspis* coincides with the base of the *P. acuminatus* Zone. The shallow marine platform deposits there are very rich in microfloras and in micro- and macrofaunas, except graptolites (see Lespérance 1981 for numerous contributions and earlier references). On the whole, the Ashgill and Llandovery acritarchs of Anticosti are very well preserved and relatively abundant, but have been described only partially (Staplin *et al.* 1965; Cramer 1970; Duffield & Legault 1981, 1982), apart from strata dated as *D. complanatus* Zone, assigned to the early or middle Ashgill (Jacobson & Achab 1985).

The Anticosti acritarchs

The quality of the palynological material at Anticosti and its age control based on shelly macrofaunas and conodonts justify a preliminary synthesis. The ranges of some taxa there are compared (Fig. 2) with those from other regions. The compilation is based on the references given in the general distribution of data (Fig. 1) and for the post-Aeronian of the same regions, following those assembled by Martin (in press; explanation of Fig. 1). This restricted choice of taxa is conditioned by personal examination of twelve samples (see Appendix) from the upper part of Member 4 of the Ellis Bay Formation, of Gamachian age, to the upper part of the Jupiter Formation, correlated with the Telychian (C_5) (Lespérance 1981). The choice could have been different, but in the present state of knowledge the comments would probably have been comparable with those below.

The observations of Duffield & Legault (1981) are confirmed with regard to the change in composition of acritarch assemblages just above the base of the Silurian as defined on the basis of the appearance of diagnostic conodonts (Barnes & McCracken 1981) within Member 7 of the Ellis Bay Formation. If the correlation proposed by Lespérance (1985) is accepted, the major change in terms of appearance of new acritarch taxa occurs within the late Gamachian, rather than in the early Llandovery. At its type locality, on the west side of Ellis Bay, the entire member, 1 to 4 m thick, is very poor in acritarchs. In particular, the locally developed biohermal bed, 1·5 to 2 m thick, above the systemic boundary is sterile. Immediately above this bed, from the base of the Becscie Formation (*sensu* Petryk 1981; sample A2B7) onwards, the majority of taxa known from other regions and of Ordovician affinities are absent. *Aremoricanium squarrosum* Loeblich & McAdam, 1971 (see synonymy in Jacobson & Achab, 1985: 171) is recognized in the early Richmondian, which is equated with latest Pusgillian to early Raw-



SELECTED ACRITARCHS FROM ANTICOSTI (QUEBEC, CANADA)	CANADA GASPÉ	U.S.A. CENTRAL	U.S.A. NE & S APPALACHIANS	BALTIC REGION & POLAND	CZECHOSLOVAKIA	ENGLAND & WALES	BELGIUM	PORTUGAL	NORTH AFRICA
DIEXALLOPHASIS REMOTA	C-T		Rh-W	W	C-H	Rh-Lu	C-D		(AeT)-D
"HOGKLINTIA DIGITATA-H. VISBYENSIS"	(RhT)T	(AeT)W	W	pC-W		W	Rh		
GONIOSPHAERIDIUM OLIGOSPINOSUM	(RhT)			pC-W					
AREMORICANIUM SQUARROSUM	(PR)	(CA)A			H				Ap.d.
ORTHOSPHAERIDIUM RECTANGULARE		A	pC-A					(CA)	
ORTHOSPHAERIDIUM INSCULPTUM	(PR)	A						Ap.d.	C?p.d.
BALTISPHAERIDIUM PLICATISPINAE				pC-A					
PHEOCLOSTERIUM SP.NOV.									
GEN. & SP. NOV. cf. RHOPALIOPHORA SP.									
MULTIPLICISPHAERIDIUM SP. 1									
TYLOTOPALLA SP.	(RhT)		A-W			Rh-W	(CA)Rh		
EUPOIKILOFUSA aff. E. AMPULLIFORMIS	Rh(RhT)								
DOMASIA SP.	T	(AeT)W	(AeT)W	TW		Ae-W	Ae-W		(WLu)pd
DILATISPHAERA WILLIEREAE	T					T-?Lu	Ae-T		

1 North American standard
2 British standard
3 I.U.G.S. standard

——— Occurrence of taxon on Anticosti
— — — — Occurrence of taxon elsewhere

pC: Ordovician, pre-Caradoc, C: Caradoc, A: Ashgill, P: Pusgillian, R: Rawtheyan, H: Hirnantian, Rh: Rhuddanian, Ae: Aeronian, T: Telychian, W: Wenlock, Lu: Ludlow, D: Devonian, p.d.: palynological dating, chronostratigraphic units in parentheses are undifferentiated, ? before chrono stratigraphic symbol: dubious record of taxon, ? after chronostratigraphic symbol: dubious age attribution.

Fig. 2 Ranges of selected Anticosti acritarchs in other regions.

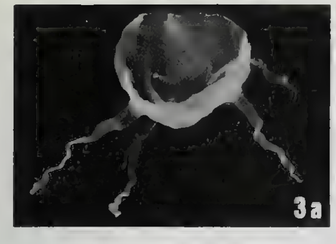
theyan by Barnes *et al.* (1981). The disappearance of *Orthosphaeridium rectangulare* (Eisenack) Eisenack, 1968 (Figs 4a, b; see synonymy in Elaouad-Debbaj 1981: 48) and of *O. insculptum* Loeblich, 1970 (Figs 3a, b) occurs within an unobserved interval in the Gamachian, between the upper parts of Member 5 (about 5 m below its top; sample A2B3) and Member 6 (0·3 m below its top; sample A2B4) of the Ellis Bay Formation. *Baltisphaeridium plicatispinae* Górka, 1969 (Fig. 9) extends, according to Duffield & Legault (1981), into Member 7, below the biohermal bed. The appearance of taxa of Silurian affinities, which occurs mainly and progressively from the base of the Becscie Formation onwards, begins in the Gamachian, no later than the upper part of Member 5 (sample A2B3), source of the present example of *Multiplicisphaeridium* sp. 1, *sensu* Duffield & Legault 1981 (Fig. 16). The latter recalls the '*M. forquiferum–M. forquillum*' group found by Cramer & Díez (1972) in the late Llandovery of Kentucky. *Eupoikilofusa* aff. *E. ampulliformis* (Figs 14a, b), which appears at the base of the Becscie Formation (sample A2B7), earliest Llandovery, is close to a Llandovery species known from the early Rhuddanian in Belgium (Martin 1974). The entry of *Domasia* Downie, 1960, emend. Hill 1974 (Fig. 6) and *Tylotopalla* Loeblich, 1970 (Fig. 10) on the one hand, and of *Dilatisphaera williereae* (Martin) Lister 1970 (Fig. 5) on the other, occurs in the Jupiter Formation at levels that are correlated (Barnes & McCracken 1981) respectively with the late Aeronian (C_1–C_2; sample A6A, about 3 m above base of Member 3) and with the Telychian (C_5; sample A7A1, 4 m below top of the Jupiter Formation). As yet no diacrodian has been identified from the upper part of the Gamachian, and no form suspected of being reworked from the Ordovician has been found in the Llandovery of Anticosti.

The richness and variety of the microfloras in the Gamachian and Llandovery at Anticosti will lead inevitably to the introduction of new taxa, some of which will be index fossils. As an example, two forms from the Ellis Bay Formation (sample A2B3) are illustrated for the first time here and left in open nomenclature: *Pheoclosterium* sp. nov. (Figs 7a, b) and Gen. et sp. nov. cf. *Rhopaliophora* (Fig. 8). The only species formally assigned to the former genus, *Pheoclosterium fuscinulaegerum* Tappan & Loeblich, 1971, is characteristic of the late Ordovician. Its range (see Jacobson & Achab 1985 for all references) is from the Edenian of Indiana (Kope Formation; Tappan & Loeblich 1971; Colbath 1979) and from the Onnian, highest Caradoc, in Shropshire, England (upper part of Onny Shales; Turner 1984) to the Hirnantian in Czechoslovakia (Kosov Formation, Vavrdová 1982). The second acritarch, cf. *Rhopaliophora*, differs from that exclusively Ordovician genus in its opening and resembles '*Hystrichosphaeridium*' *wimani*

Figs 3–16 Acritarchs from Anticosti. All figured specimens are in the type fossil collection of the Geological Survey of Canada, Ottawa, and have numbers with the prefix GSC.

Figs 3, 4, 7–9, 12, 15, 16: sample A2B3, Ellis Bay; Ellis Bay Formation, upper part of Member 5, Gamachian. Figs 11, 13, 14: sample A2B7, Ellis Bay; lowermost Becscie Formation, Llandovery, correlated with Rhuddanian, A_{2-4}. Figs 5, 6, 10: sample A7A1, 4 km southeast of Pointe Sud-Ouest; upper part of Jupiter Formation, Llandovery, correlated with Telychian, C_5. Age assignments according to Lespérance (1981).

Fig. 3 *Orthosphaeridium insculptum* Loeblich 1970. GSC 82877. Fig. 3a, × 400; Fig. 3b, enlargement, × 3000, of base of left process. **Fig. 4** *Orthosphaeridium rectangulare* (Eisenack) Eisenack 1968. GSC 82878. Fig. 4a, enlargement, × 2000, of base of left lower process. Fig. 4b, × 200. **Fig. 5** *Dilatisphaera williereae* (Martin) Lister 1970. GSC 82879, × 1000. **Fig. 6** *Domasia limaciformis* (Stockmans & Willière) Cramer 1970. GSC 82880, × 500. **Fig. 7** *Pheoclosterium* sp. nov. GSC 82881. Fig. 7a, enlargement, × 3000, of upper median processes. Fig. 7b, × 750. **Fig. 8** Gen. et sp. nov. cf. *Rhopaliophora* sp. GSC 82882, × 300. **Fig. 9** *Baltisphaeridium plicatispinae* Górka 1969. GSC 82883, × 300. **Fig. 10** *Tylotopalla* sp. GSC 82884, × 750. **Figs 11, 12** '*Hogklintia digitata–H. visbyensis*'. Fig. 11, GSC 82885, × 250. Fig. 12, GSC 82886, × 100. **Fig. 13** *Goniosphaeridium oligospinosum* (Eisenack) Eisenack 1969. GSC 82887, × 250. **Fig. 14** *Eupoikilofusa* aff. *E. ampulliformis*, *sensu* Duffield & Legault, 1981. GSC 82888. Fig. 14a, × 1000; Fig. 14b, enlargement, × 5000, of lower right part of vesicle. **Fig. 15** *Diexallophasis remota* (Deunff) Playford 1977. GSC 82889, × 500. **Fig. 16** *Multiplicisphaeridium* sp. I, *sensu* Duffield & Legault 1981. GSC 82890, × 500.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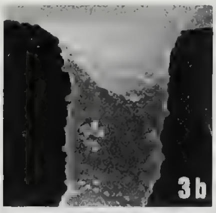
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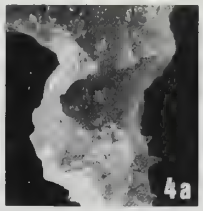
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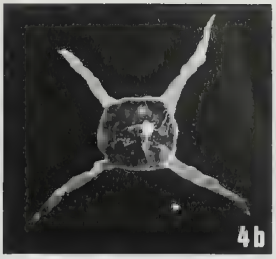
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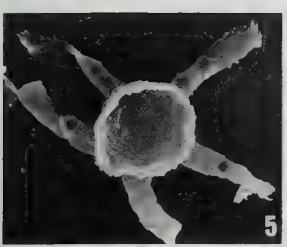
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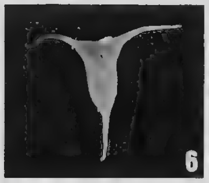
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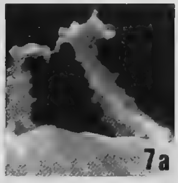
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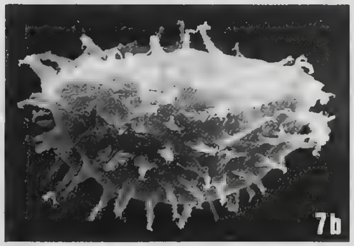
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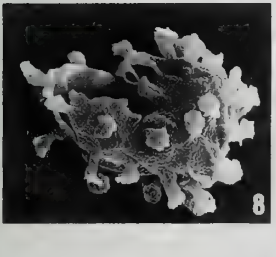
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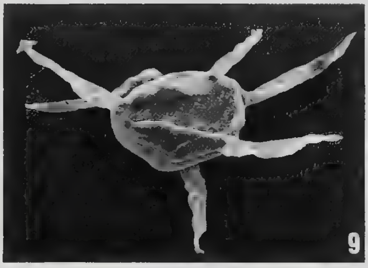
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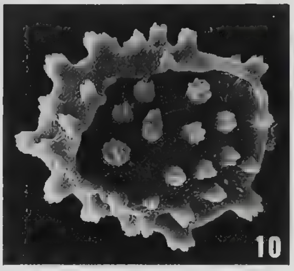
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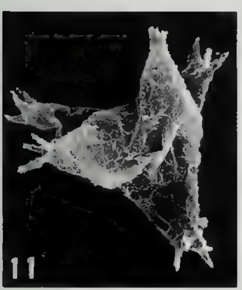
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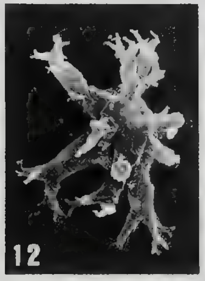
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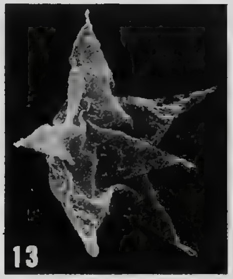
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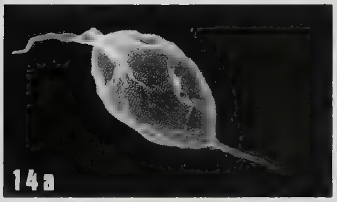
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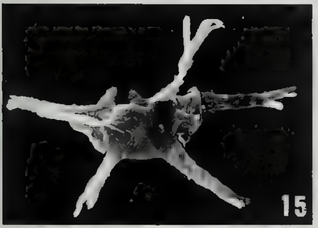
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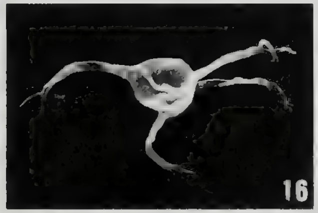
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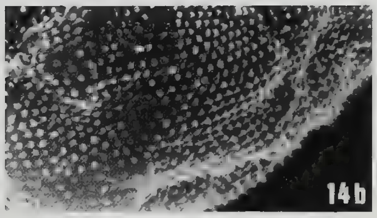
14b

Eisenack, 1968, determined by its author from the latest Ashgill of Gotland (Bornholmer Stufe F2 from an erratic boulder at Oil Myr).

On Anticosti, in both the Ashgill and the Llandovery, there are geographically widespread forms with long stratigraphical ranges that are difficult to define because of their wide, continuous morphological variability within a single sample; examples are *Diexallophasis remota* (Deunff) Playford 1977 (Fig. 15) and the '*Hogklintia digitata–H. visbyensis*' complex (Figs 11, 12). The recurrent abundance in certain Ordovician and Silurian strata, notably on Anticosti and in the Baltic region, of the latter complex and of, for instance, *Goniosphaeridium oligospinosum* (Eisenack) Eisenack 1969 (Fig. 13) probably results from particular palaeoenvironmental conditions; the latter led Cramer & Díez (see 1974 for earlier references) to postulate a certain degree of provincialism linked to palaeolatitudes for Silurian acritarchs.

Acritarch data for the latest Ordovician and earliest Silurian are as yet too disparate to permit reliable palaeogeographic reconstructions. Data from Anticosti indicate affinities and possibilities for correlation as follows. The Gamachian microfloras contain taxa known from the late Ordovician of central U.S.A. and/or the pre-Hirnantian Ashgill of Gaspé, and from the Ordovician of Europe (Baltic region and Portugal) and North Africa (Libya). In particular, the evolutionary scheme proposed by Loeblich & Tappan (1971) for the genus *Orthosphaeridium* Eisenack 1968, notably in part of the Cincinnatian of central U.S.A. and in the late Ashgill of the Baltic region, Gotland and Estonia, may be applied to the Gamachian of Anticosti and the late Ordovician of Portugal. The possibilities for correlation offered by the Llandovery acritarchs of Anticosti concern affinities with, principally and in decreasing order, the Gaspé area of Canada, England and Wales, Belgium and the U.S.A. In particular, the first occurrences of *Domasia* and of *Dilatisphaera williereae*, the levels of which are still inadequately known on Anticosti, should permit correlation with at least the Aeronian and the Telychian of the Anglo-Welsh area. Palynological data for the Rhuddanian of the latter area allow only a local zonation at present.

Conclusions

Owing to the dearth of published data, acritarchs have not been used directly as one of the criteria for the choice of an Ordovician–Silurian boundary during the activities of the I.U.G.S. working group from 1974 to 1985. The Anticosti deposits are those likely to provide the most reliable palynological correlations, not only in the immediate vicinity of the systemic boundary but also at least from the early to middle Ashgill to the late Llandovery (Telychian, C_5). This view is supported by the indication both of relatively continuous data and of direct correlations with the Gaspé area from the base of the Rhuddanian upwards, and the Anglo-Welsh area from the Aeronian upwards.

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Appendix

Locality data for Anticosti Island, Province of Québec, Canada. All locality numbers in Lespérance (1981: 1).

Loc. A-2A: Pointe Laframboise area. Sample A-2A1: Ellis Bay Formation, Member 7, 0·40 m above oncolithic bed. Sample A-2A2: Becscie Formation, 0·60 m above base.

Loc. A-2B: west side of Ellis Bay, section proposed as Ordovician–Silurian boundary stratotype by Barnes & McCracken (1981). Samples A-2B2 to A-2B6: Ellis Bay Formation; A-2B2: member 4, 3 m below top of member; A-2B3: member 5, 5 m below top of member; A-2B4: member 6, 0·30 m below top of member; A-2B5 and A-2B6: member 7, respectively just above and 0·75 m above oncolithic bed. Samples A-2B7 to A-2B9: Becscie Formation. A-2B7: immediately above the biohermal level of member 7 of the Ellis Bay Formation. A-2B8 and A-2B9: respectively 1·30 m and 25 m (approximately) above A-2B7.

Loc. A-6A and sample A-6A: Cap Jupiter, north of mouth of Rivière Jupiter, Jupiter Formation, about 3 m above base of member 3.

Loc. A-7A: 4 km southeast of Pointe du Sud-Ouest. Sample A-7A1: Jupiter Formation, 4 m below its top.

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Brachiopods across the Ordovician–Silurian boundary

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Synopsis

Most of the late Ordovician brachiopod superfamilies also extend into the early Silurian, although the Gonambonitacea become extinct at or near the Ordovician–Silurian boundary and the earliest Cyrtiacea are found very close above it. Faunas close to the boundary are reviewed and listed, and the Hirnantian faunas of the latest Ordovician are found to be richer than the earliest Silurian Rhuddanian faunas in both abundance and diversity.

Introduction

At the time the Treatise on Invertebrate Paleontology brachiopod volume (Williams *et al.* 1965) was written, 44 brachiopod genera were recorded with ranges spanning the Ordovician–Silurian boundary, and in addition there were various families and subfamilies whose ranges spanned the boundary even if the recorded ranges of individual genera within them did not. The superfamilies involved are the Lingulacea, Trimerellacea, Discinacea, Craniacea, Orthacea, Enteletacea, Tripleciacea, Eichwaldiacea, Plectambonitacea, Strophomenacea, Davidsoniacea, Chonetacea, Porambonitacea, Pentameracea, Rhynchonellacea, Atrypacea and Athyridacea—a list which in itself demonstrates the morphological variability and diversity of the phylum in Ordovician–Silurian boundary times.

However, rather than review each family, genus or species in turn here, it is more relevant to consider the brachiopod faunas actually recovered from strata near the boundary. In general the middle Ashgill was a period of great diversity among the brachiopods, but this diversity was reduced when the Rawtheyan endemic faunas, for example of North America (the late Richmondian) and Europe (e.g. the Boda Limestone of Sweden) gave way to the more cosmopolitan, and hence in total less diverse, faunas of Hirnantian times. Similarly, the profound effect of the Ordovician–Silurian boundary glacial episode made the subsequent recovery and build-up of the brachiopod faunas rather slow, and thus, even where the earliest Llandovery time is represented by rock (and not by the usual unconformity), the numbers and more particularly the diversity of the brachiopod faunas were rather poor.

Latest Ordovician and earliest Silurian brachiopods

In the following lists the records are reproduced of reliable determinations from relatively recent papers on brachiopods of Hirnantian and early Rhuddanian ages respectively. In most cases they are as the original authors determined them, but with 'aff.' or 'cf.' omitted, and sometimes with genera or species updated by subsequent works. They are from the following authors and localities: A, uppermost Ellis Bay and lowermost Becscie Formations, Anticosti Island, Canada (Cocks & Copper 1981); B, Kosov Formation, Bohemia, Czechoslovakia (Marek & Havlíček 1967; Havlíček 1977); D, Durben Horizon, Kazakhstan, USSR (Nikitin *et al.* 1980); E, Lower Edgewood Group, Oklahoma, USA (Amsden 1974); G, High Mains Sandstone and Lady Burn Formation, Girvan, Scotland (Cocks & Toghill 1973; Harper, this volume); H, St Martin's Cemetery Horizon, Haverfordwest, Wales (Cocks & Price 1975); I, Hol Beck, England (Temple 1965); K, Kildare, Ireland (Wright 1968); L, Bronydd Formation, Llandovery, Wales (Cocks *et al.* 1984); M, *persculptus* and *acuminatus* Zones, Mirny Creek, north-east USSR (Koren *et al.* 1983); O, Langøyene and Langåra Formations, Oslo–Asker district, Norway (Brenchley & Cocks 1982; Cocks 1982) and Myren Member (Baarli & Harper 1986); P, Unit 5, White Head Formation, Percé, Québec, Canada (Lespérance & Sheehan 1976,

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1981); R, Varbola Formation, Estonia, USSR (Rubel 1970); S, Stawy, Poland (Temple 1965); V, Dalmanitina Beds, Våstergotland, Sweden (Bergström 1968); W, Hirnant Beds, Wales (Temple 1965); X, Hirnantian Beds, Keisley, England (Temple 1968); Y, Kuanyinchiao Beds, Yichang, China (Rong 1984*a*); Z, Artchalyk and Minkutchar Beds, Zeravshano–Gissar section, Altai Mountains, USSR (Nikiforova 1978).

The latest Ordovician (Hirnantian) records from these localities are as follows:

Lingulacea: *Lingula* sp. H, O; *Lingulella* sp. I, S; *Palaeoglossa* sp. V; *Craniops/Paracraniops* sp. H, O, V, X.

Discinacea: *Trematis norvegica* Cocks O; *Orbiculoidea concentrica* (Wahlenberg) H, V, S; *Orbiculoidea* sp. O.

Craniacea: *Acanthocrania* sp. O, X; *Philhedra grayii* (Davidson) X; *Philhedra* sp. H, V; *Philhedra*? *stawyensis* Temple I, S; *Philhedrella cribrum* Temple X; *Philhedrella* sp. A, O.

Orthacea: *Comatopoma sororia* Marek & Havlíček B; *Comatopoma* sp. O; *Dolerorthis intermedius* Nikiforova M; *Dolerorthis praeclara* Temple X; *Dolerorthis savagei* Amsden E; *Dolerorthis* sp. O; *Geraldibella bella* (Bergström) M, V; *Geraldibella giraldi* (Bancroft) H; *Giraldibella subsilurica* (Marek & Havlíček) B; *Glyptorthis* sp. G, O; *Hesperorthis* sp. M, O; *Nicolella* sp. O; *Orthostrophella* sp. E; *Plaesiomys* sp. G; *Platystrophia* sp. E, G, O; *Skenidioides scoliodus* Temple X; *Skenidioides* sp. H, O; *Toxorthis mirabilis* Rong Y; *Toxorthis proteus* Temple X.

Enteletacea: *Dalmanella biconvexa* Williams H; *Dalmanella cicatrica* Nikitin D; *Dalmanella edgewoodensis* Savage E; *Dalmanella pectinoides* Bergström B, V; *Dalmanella testudinaria* (Dalman) A, B, H, I, K, M, O, P, S, V, W, Y; *Dicoelosia* sp. E, X; *Diceromyonia*? *sera* Amsden E; *Draborthis caelebs* Marek & Havlíček B, V, X, Y; *Drabovia agnata* Marek & Havlíček B; *Drabovia westrogothica* Bergström V; *Drabovia* sp. O, X; *Dysprosorthis sinensis* Rong Y; *Epitomyonia* sp. O; *Hirnantia noixella* Amsden E; *Hirnantia sagittifera* (M'Coy) B, D, G, H, I, K, M, O, P, S, V, W, X, Y; *Hirnantia* sp. A; *Horderleyella bouceki* (Havlíček) S, W; *Horderleyella fragilis* Bergström V; *Isorthis* sp. M; *Kinnella kielanae* (Temple) B, P, S, V, W, X, Y; *Leptoskelidion loci* Cocks O; *Leptoskelidion septulosum* Amsden E; *Mendacella*? sp. E; *Mirorthis mira* Zeng Y; *Onniella kalvoya* Cocks O; *Onniella*? *yichangensis* Zeng Y; *Paucicrura* sp. O; '*Pionodema*' *retusa* Temple X; *Ravozetina rava* Marek & Havlíček B; *Reuschella inexpectata* Temple X; *Trucizetina subrotundata* Havlíček B; *Trucizetina yichangensis* Zeng Y; *Visbyella*? sp. [= *Kayserella* sp. nov. of Temple] X.

Gonambonitacea: *Kullervo*? sp. O.

Tripleciacea: *Cliftonia psittacina* (Dalman) B, H, K, O, V; *Cliftonia obovata* Chang Y; *Cliftonia tubulistriata* (Savage) E; *Cliftonia* sp. D, M; *Onychoplecia* sp. X, Y; *Oxoplecia* sp. O; *Triplesia protea* Oradovskaya M; *Triplesia sanxiaensis* Zeng Y; *Triplesia* sp. O.

Plectambonitacea: *Aegiromena convexa* Chang Y; *Aegiromena durbenensis* Nikitin D; *Aegiromena ultima* Marek & Havlíček B, Y; *Aegiromena* sp. X; *Anisopleurella novemcostata* Nikitin D; *Chonetoidea papillosa* (Reed) H; *Eochonetes* sp. G; *Eoplectodonta nesnakomkaensis* Oradovskaya M; *Eoplectodonta rhombica* (M'Coy) O; *Eoplectodonta oscitanda* Cocks O; *Eoplectodonta* sp. D; *Leangella cylindrica* (Reed) O, V; *Rugosowerbyella ambigua* (Reed) D; *Sampo* sp. O; *Sericoidea*? O.

Strophomenacea: *Aphanomena parvicostellata* Rong Y; *Aphanomena schmalenseei* Bergström V; *Biparetis paucirugosus* Amsden M; *Eopholidostrophia* sp. G; *Eostropheodonta bublitschenki* Nikitin D; *Eostropheodonta hirnantensis* (M'Coy) including *E. lucavica* and *E. siluriana* A, B, G, I, K, M, O, P, S, V, W; *Eostropheodonta whittingtoni* Bancroft H; *Katastrophomena* sp. O; *Kjaerina*? sp. O; *Kjerulfina*? sp. V; *Leptaena aequalis* Amsden M; *Leptaena martinensis* Cocks H; *Leptaena rugosa* Dalman B, D, V; *Leptaena* sp. E, O; *Leptaenopoma trifidum* Marek & Havlíček B, D, K, V, Y; *Paromalomena polonica* (Temple) B, D, I, S, X, Y; *Rafinesquina*? *latisculptilis* (Savage) E, M; *Rafinesquina stropheodontoides* (Savage) E; *Rafinesquina ultrix* Marek & Havlíček B, D; *Rafinesquina urbicola* Marek & Havlíček B, D; *Titanomena grandis* Bergström V.

Davidsoniacea: *Coolinia convexa* (Savage) E; *Coolinia dalmani* Bergström A, O, V; *Coolinia propinqua* (Meek & Worthen) E; *Coolinia* sp. M, Y; *Fardenia comes* Marek & Havlíček B; *Fardenia* sp. G, X.

Porambonitacea: *Parastrophinella gracilis* Oradovskaya M; *Parastrophina* sp. O.

Pentameracea: *Brevilamnulella kjerulfi* (Kjaer) O; *Brevilamnulella thebesensis* (Savage) E, M; *Brevilamnulella undatiformis* Rozman M; *Holorhynchus giganteus* Kjaer O; *Tcherskidium unicum* (Nikolaev) M.

Rhynchonellacea: *Dorytreta* sp. Y; *Hypsiptycha* sp. G; *Rostricellula* sp. G, O; *Rhynchotrema*? sp. M; *Stegerhynchus concinna* (Savage) E, M; *Stegerhynchus*? sp. E, O; *Thebesia admiranda* Oradovskaya M; *Thebesia scopulosa* Cocks O; *Thebesia thebesensis* (Foerste) E.

Atrypacea: *Eospirigerina gaspeensis* (Cooper) M; *Eospirigerina prisca* Oradovskaya M; *Eospirigerina putilla* (Hall & Clarke) E; *Eospirigerina sublevis* Rozman M; *Eospirigerina* sp. G, O; '*Homoeospira*' *fiscellostriata* Savage E; *Plectatrypa* sp. M; *Protatrypa* sp. X; *Protozyga gastrodes* Temple X; *Zygospira fallax* Marek & Havlíček B; *Zygospira* sp. O.

Athyracea: *Cryptothyrella crassa* (J. de C. Sowerby) *incipiens* Williams G, H, K, Y; *Cryptothyrella ovoides* (Savage) E; *Cryptothyrella terebratulina* (Wahlenberg) M; *Cryptothyrella* sp. B, X; *Hindella cassidea* (Dalman) O, ?P, ?A; *Hyattidina* sp. M; *Plectothyrella crassicostis* (Dalman) [ex *platystrophoides* Temple] B, I, K, P, S, V, W, Y; *Plectothyrella*? *mirnyensis* Oradovskaya M.

Eichwaldiacea: *Dictyonella* sp. E.

The earliest Silurian (lower part of the Rhuddanian) records from these localities are:

Lingulacea: *Lingula* sp. G.

Discinacea: *Orbiculoidea* sp. H.

Orthacea: *Dolerothis plicata* (J. de C. Sowerby) O; *Dolerorthis sowerbyiana* (Davidson) L; *Dolerorthis* sp. O, R; *Giraldiella* sp. L, Z; *Hesperorthis imbecilla* Rubel R; *Platystrophia* sp. R; *Schizonema* sp. L, O; *Ptychopleurella* sp. R; *Skenidioides scoliodus* Temple M; *Skenidioides woodlandensis* Reed O; *Skenidioides* sp. H, L, O.

Enteletacea: *Dalejina* sp. R; *Dicoelosia osloensis* Wright O; *Dicoelosia* sp. L; *Draborthis*? sp. M; *Epitomyonia* sp. O; *Fascifera* sp. O; *Howellites* sp. O; *Isorthis neocrassa* Nikiforova Z; *Isorthis prima* Walmsley O; *Isorthis* sp. A; *Kinnella* sp. O; *Onniella mediocra* Rubel R; *Ravozetina* sp. L, O; *Resserella* sp. H, L; *Reuschella* sp. O; *Visbyella* sp. L.

Tripleciacea: *Triplesia* sp. L, O.

Plectambonitacea: *Aegiria norvegica* Öpik O; *Anisopleurella* sp. L; *Anisopleurella gracilis* (Jones) H; *Eoplectodonta duplicata* (J. de C. Sowerby) L, O; *Eoplectodonta* sp. H; *Leangella scissa* (Davidson) L, O.

Strophomenacea: *Eopholidostrophia* sp. A, L; *Eostropheodonta* sp. H; *Furcitella* sp. L; *Katastrophomena* sp. L; *Leptaena aequalis* Amsden M; *Leptaena contermina* Cocks A; *Leptaena haverfordensis* Bancroft O; *Leptaena reedi* Cocks L, O; *Leptaena valentia* Cocks L; *Leptaena* sp. H, O, R; *Leptostrophia reedi* (Bancroft) A; *Leptostrophia* sp. L.

Davidsoniacea: *Fardenia* sp. G, L, R.

Porambonitacea: *Parastrophinella* sp. Z.

Pentameracea: *Clorinda malmoyensis* St Joseph Z; *Clorinda undata* (J. de C. Sowerby) H, L, O; *Clorinda* sp. R; *Stricklandia lens* (J. de C. Sowerby) H, L, O, R, Z; *Virgiana* sp. Z; *Virgianella sogdianica* Nikiforova & Sapelnikov Z.

Rhynchonellacea: *Rhynchotrema* sp. L; *Rhynchotreta*? sp. G.

Atrypacea: *Alispira gracilis* Nikiforova R; *Clintonella aprinis* (Verneuil) R; *Clintonella* sp. R; *Eospirigerina porkuniana* Rubel R; *Idiospira* sp. O; *Eospirigerina* sp. H, O, Z; *Meifodia recta alia* Nikiforova Z; *Meifodia* sp. L, O; *Plectatrypa imbricata* (J. de C. Sowerby) Z; *Plectatrypa* sp. L; *Protatrypa malmoeyensis* Boucot, Johnson & Staton O, Z; *Protatrypa* sp. M; *Protozyga* sp. L; *Zygospiraella* sp. M, Z; *Zygospiraella duboisi* (Verneuil) R.

Athyracea: '*Atrypina*' *gamachiana* Twenhofel A; *Cryptothyrella angustifrons* (Salter) L, G; *Cryptothyrella crassa* (J. de C. Sowerby) H, L; *Cryptothyrella* sp. A, R; '*Hindella*' *extenuata* Rubel R; *Hyattidina* sp. M.

From these lists it can be seen that the cited faunas carried 90 genera in the Hirnantian and only 54 in the early Rhuddanian, with 32 genera in common between the two lists. Part of this numerical discrepancy can be explained by the greater number of faunal lists available for beds of Hirnantian age (18), compared with only 8 for the early Rhuddanian; nevertheless that discrepancy can itself be explained by the fewer number of early Llandovery age faunas that can actually be found. Moreover, whereas the Hirnantian faunas can often be found in abundance (for example in China—Rong 1984*a*, *b*), the early Rhuddanian faunas are often very sparse both in numbers and diversity, and also in the actual size of the specimens, all of which explains why monographic treatment of them has been rather neglected, particularly by comparison with the much richer and more diverse later Rhuddanian faunas, which are relatively well described (e.g. Temple 1970). In addition, presumably because of the glacially-caused eustatic lowering of sea level which peaked during the Hirnantian, there are many sections in which only the Hirnantian is represented by shelly deposits and with the beds above and below in which the only macrofossils are graptolites.

Missing from both of the above lists are representatives of the Trimerellacea, Acrotretacea, Siphonotretacea and Chonetacea, all of which have reliable records from both late Ordovician and early Silurian rocks, but not from beds very close to the boundary; and from the early

Rhuddanian list the Craniacea and the Eichwaldiacea, which also yield representatives from later horizons in the Llandovery. The only brachiopod superfamily which appears to have become extinct at the end of the Hirnantian is the Gonambonitacea (although a few lower taxa such as the Trematidae also disappeared then); and the only new superfamily to appear anywhere near the base of the Silurian is the Cyrtiacea, whose earliest records, although not accurately dated in detail, come from beds in Tasmania extremely close to the boundary (Sheehan & Baillie 1981). In general, however, the degree of extinction across the boundary appears to have been far less than previously reported, largely because earlier studies have not concentrated on latest Ashgill and earliest Llandovery rocks. The extinctions at the end of the Hirnantian do not appear to have been greater than at the end Caradoc or end Rawtheyan. This is exemplified by a recent review of the atrypoids by Copper (1986), who states that only two genera, *Idiospira* and *Cyclospira*, may have become extinct near the boundary, and even these two have been reported (e.g. Baarli & Harper 1986) from early Silurian rocks. The strong 'Silurian' elements in the spire-bearer fauna, for example *Hindella* and *Eospirigerina*, actually appeared in late Rawtheyan times.

Unfortunately no evolutionary gradation within a single genus has been adequately studied across the boundary, and thus no perfect recognition of the boundary by brachiopods is yet possible. The most striking changes in closely related groups are seen in the Pentameracea, which can be found in virtually rock-building abundance in some beds both above and below the boundary, although only rarely in the earliest Rhuddanian. In the Hirnantian, *Holorhynchus*, *Brevilamnulella*, and others dominate the fauna, whereas in the Rhuddanian their place is taken by *Stricklandia*, *Clorinda*, and a wide diversity of genera in the then tropical areas of the USSR (Nikolaev *et al.* 1977) and, rather later, *Virgiana* and *Platymerella* in the USA. In the east Baltic, *Borealis* is known from as low as the *vesiculosus* Zone (Boucot *et al.* 1969).

The exact age, in terms of graptolite zones, of the various brachiopod faunas from near the systemic boundary, in particular the *Hirnantia* fauna, is also of great relevance in international correlation. In continuous sections, most *Hirnantia* faunas underlie beds bearing *persculptus* Zone graptolites, for example in the vast outcrop area in China, and in general the fauna is undoubtedly of *extraordinarius* Zone age or older; it spans four graptolite zones in China (Rong 1984*b*). However, in at least two places it occurs in beds with and above *persculptus* Zone graptolites. One is in Kazakhstan, USSR (Apollonov *et al.*, this volume, p. 145), and the other is in the Lake District, England, where Locality 74/1 of Hutt (1974: 15) in Yewdale Beck, Cumbria (National Grid ref. SD 3073 9858) has yielded to J. E. Hutt (registered numbers BC 7217–7236), in order of abundance, *Kinnella kielanae* (Temple), *Mirorthis mira* Zeng, *Plectothyrella crassicostis* (Dalman), *Cyclospira* sp., *Hirnantia* sp. and other indeterminate orthids and dalmanellids, identified by the author and Rong Jia-yu. In addition the same bed has yielded many graptolites (J. E. Hutt, pers. comm. 1986), including *Climacograptus medius* Törnquist, *C. normalis* Lapworth, *C. miserabilis* Elles & Wood, *Glyptograptus persculptus* (Salter), *Diplograptus* ex gr. *modestus* Lapworth and *Monograptus ceryx* Rickards & Hutt. These new records endorse the most preferable systemic boundary at the base of the *acuminatus* Zone.

Acknowledgements

I am most grateful to Rong Jia-yu and A. J. Boucot, who helpfully commented on the first draft of this paper.

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Chitinozoan stratigraphy in the Ashgill and Llandovery

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Synopsis

There is little published information on chitinozoan faunas from sections with continuous sedimentation across the Ordovician–Silurian boundary. Most boundary sections from which chitinozoans have been described include a hiatus. To aid determination of the extent of any hiatus, the ranges of thirty-one diagnostic chitinozoan species from Ashgill and Llandovery strata are documented, with reference to the British standard graptolite zonation. The composition of the chitinozoan faunas at the Ordovician–Silurian boundary is discussed, and the influence of ecological factors is assessed.

Introduction

Chitinozoans are organic-walled microfossils known from marine sedimentary rocks of Ordovician to Devonian age. Although pertinent information is missing from, for instance, Australia and the East Indies, it is no exaggeration to say that chitinozoans have great stratigraphical potential on a world-wide basis. However, our knowledge of the chitinozoan faunas at the Ordovician–Silurian boundary is still scanty. Chitinozoans from sections with continuous sedimentation across the boundary are known only from Anticosti Island, Québec (Achab 1981), Skåne, Sweden (Grahn 1978) and probably Estonia (Nestor 1976, 1980*a*, 1980*b*, and personal communication 1985; Nõlvak 1980, and personal communication 1985). Faunas from sections with a small hiatus have been described from Libya (Molyneux & Paris 1985; Hill *et al.* 1985), the Cincinnati Region, midcontinent U.S.A. (Grahn 1985; Grahn & Bergström 1985; M. A. Miller, personal communication 1985) and the Brabant Massif, Belgium (Martin 1973). References to other papers with relevant data on Ashgill and/or Llandovery chitinozoan faunas will be made in context. To help in determining the extent of any hiatus, the ranges of selected chitinozoan species from the early Ashgill to the late Llandovery are documented here. The total range of each species (Fig. 1) is defined according to the British standards for the Ashgill (*sensu* Williams 1983) and Llandovery (*sensu* Cocks *et al.* 1984) Series.

Diagnostic Ashgill chitinozoans

Many Ashgill chitinozoan species are long-ranging and persist from the middle or lower Ordovician. Only a few species are restricted to the Ashgill (Fig. 1). The chitinozoans can be divided into a pre-Hirnantian fauna, and a fauna that ranges into the Hirnantian (Figs 2–12). Ashgill chitinozoans from Great Britain are virtually unknown. The type Hirnantian is barren of chitinozoans. In older strata, *Tanuchitina bergstroemi* occurs in the Rawtheyan (F. Paris, personal communication 1985).

In North Africa *Armoricochitina nigerica* and *Calpichitina*? *lenticularis* are very characteristic for the late Ashgill (Elaouad-Debbaj 1984; Molyneux & Paris 1985; J. C. Jaglin, personal communication 1985). These species are also known from SW Europe (Paris 1981). *Acanthochitina*? *rashidi*, *Ancyrochitina merga*, *Plectochitina sylvanica*, and *Sphaerochitina lepta* characterize the Ashgill in midcontinent U.S.A. (Jenkins 1970; M. A. Miller, personal communication 1986, own observations). *Ancyrochitina merga* has a more spinose ornament in the lower Ashgill than higher, and all specimens of *Sphaerochitina lepta* are smooth in the mid-Ashgill but are joined by spinose forms in the upper Ashgill. No Hirnantian chitinozoans are known from the midcontinent U.S.A.

So far, *Conochitina gamachiana* has only been reported from the upper Ashgill strata of Anticosti Island (Achab 1978). Other typical associated species are *Ancyrochitina longispina*,

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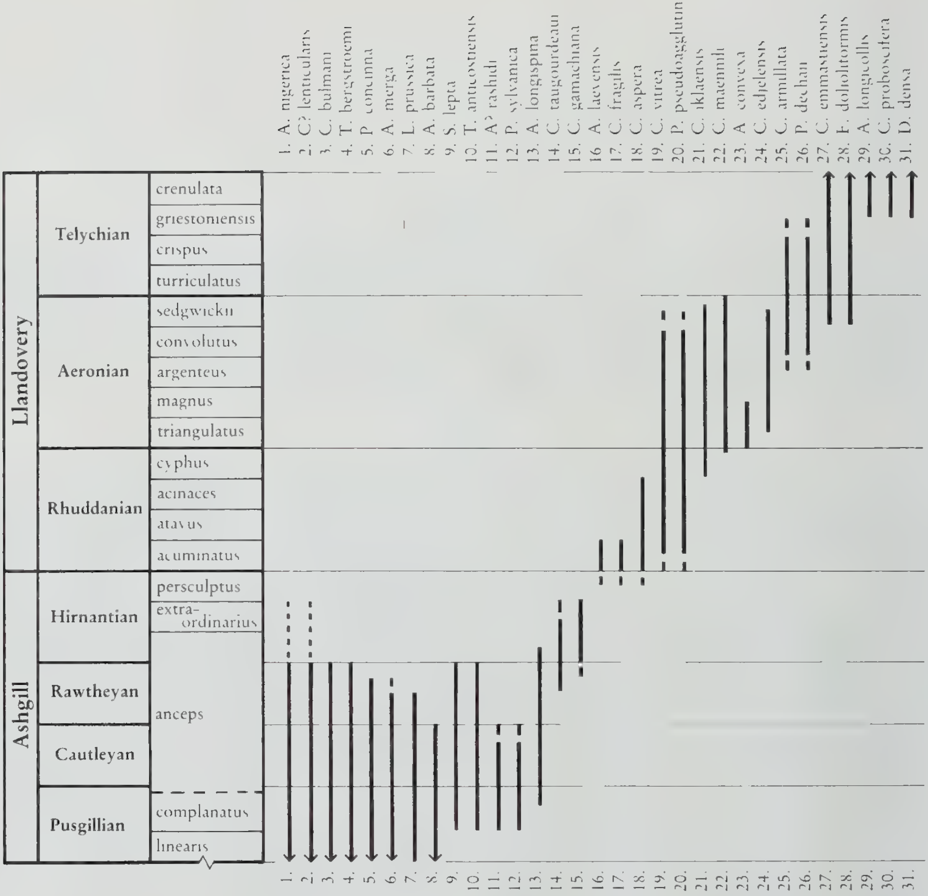

Fig. 1 Biostratigraphical ranges of selected chitinozoans from the Ashgill and Llandovery.

Figs 2–12 Selected Ashgill Chitinozoa. 2, *Tanuchitina anticostiensis*, Vauréal Formation (early Ashgill), boring RH5, Anticosti Island, Canada; SEM × 150. 3, *Calpichitina*? *lenticularis*, middle Caradoc–early Ashgill, boring E1-81 (791 m), Libya; SEM × 230. 4, *Armoricochitina nigerica*, late Caradoc–Ashgill, boring E1-81 (785–792 m), Libya; SEM × 190. 5, *Plectochitina sylvanica*, early Ashgill, boring J1-81A (3985–4000 m), Libya; SEM × 310. 6, *Sphaerochitina lepta*, Sylvan Shale (early Ashgill), Arbuckle Mountains, Oklahoma, U.S.A.; SEM × 345. 7, *Conochitina gamachiana*, Ellis Bay Formation (late Ashgill), boring A425, Anticosti Island, Canada; SEM × 270. 8, *Lagenochitina prussica*, Vormsi Stage (early Ashgill), Gotska Sandön boring (93·30–93·35 m), Sweden; SEM × 360. 9, *Tanuchitina bergstroemi*, erratic of late Caradoc age, Öland, Sweden; SEM × 130. 10, *Coronochitina taugourdeaui*, Porkuni Stage (late Ashgill), Taagepera boring, Estonia; SEM × 225. 11, *Acanthochitina barbata*, Vormsi Stage (early Ashgill), Gotska Sandön boring (96·34–96·40 m), Sweden; SEM × 140. 12, *Plectochitina concinna*, Vauréal Formation (early Ashgill), boring AF6, Anticosti Island, Canada; SEM × 310.

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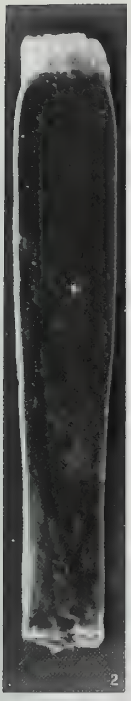
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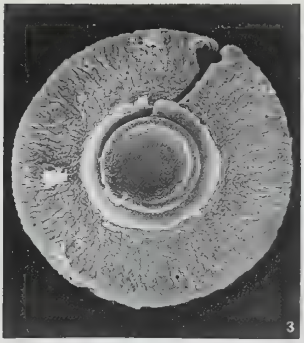
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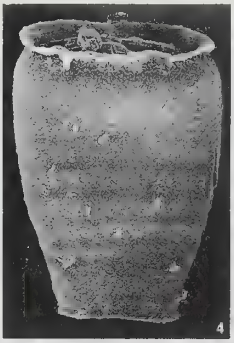
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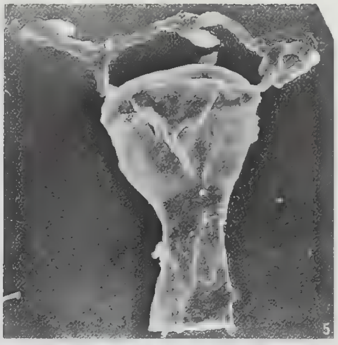
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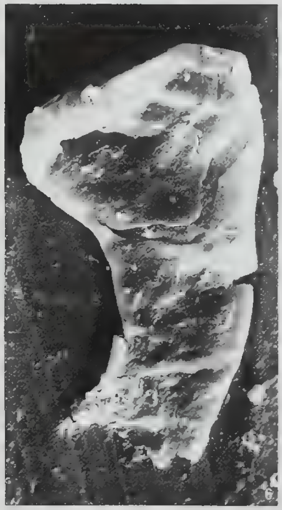
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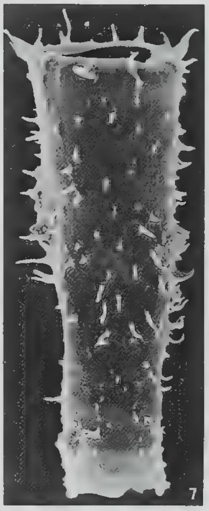
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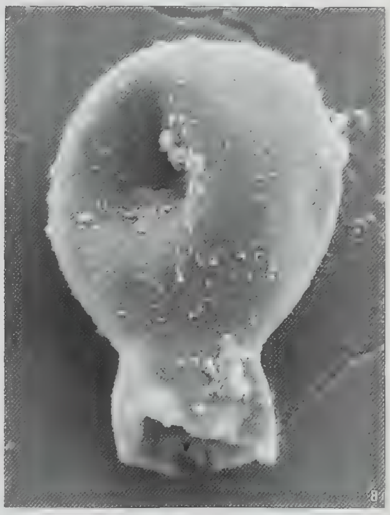
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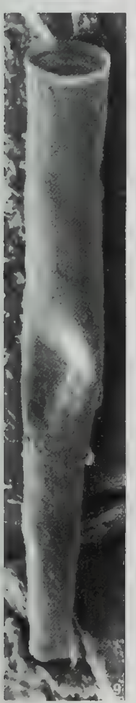
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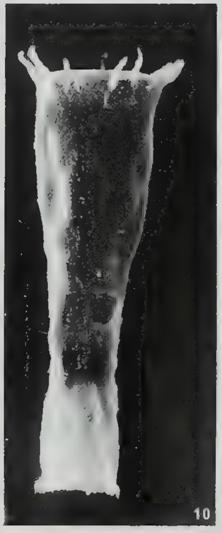
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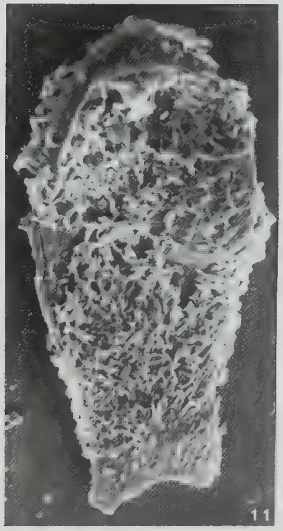
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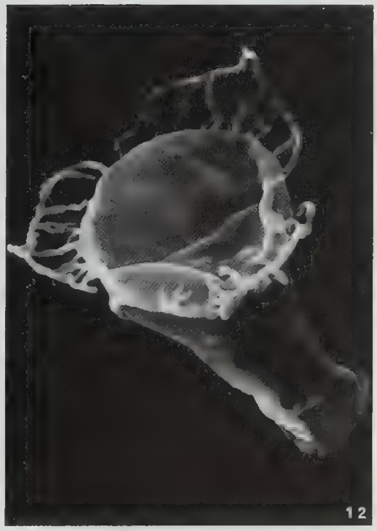
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Coronochitina bulmani, *Plectochitina concinna*, and *Tanuchitina anticostiensis* (Achab 1978). In contrast to other areas, *Hercochitina* species are very common in the Ashgill of Anticosti Island and the midcontinent U.S.A.

Acanthochitina barbata is restricted to the upper *Pleurograptus linearis* Zone in Baltoscandia (Nõlvak 1980), but has a slightly longer range in north Africa and North America. *Coronchitina taugourdeaui* is another excellent index fossil. It is one of the few chitinozoan species indicative of the Hirnantian and is known from Baltoscandia and Anticosti Island (Eisenack 1968; Nõlvak 1980; Achab 1981). *Lagenochitina prussica* and *Tanuchitina bergstroemi* are Baltoscandian species (Grahn 1982); the former is also known from the Ashgill in north Africa (Elaouad-Debbaj 1984; Molyneux & Paris 1985), the Brabant Massif, Belgium (Martin 1973; own observations) and Podolia, U.S.S.R. (Laufeld 1971).

Chitinozoan faunas at the Ordovician–Silurian boundary

Chitinozoan faunas at the Ordovician–Silurian boundary are characterized by a complex of *Ancyrochitina* (e.g. *A. ancyrea*, *A. spongiosa*) and *Cyathochitina* species (e.g. *C. campanulaeformis*, *C. kuckersiana*). Nestor (1980*a*) described *Conochitina postrobusta* from the Juuru Stage in Estonia. However, in Skåne, Sweden (Grahn 1978) and the Brabant Massif, Belgium (Martin 1973; own observations) there is no difference between late Ashgill and Llandovery specimens of *Conochitina robusta*. It is therefore uncertain whether *C. postrobusta* can be separated from Ashgill specimens of *C. robusta*.

Some Ordovician genera (e.g. *Acanthochitina*, *Hercochitina*) and typical Ordovician species (e.g. *Desmochitina* gr. *minor*, *Conochitina* gr. *micracantha*) disappear in the top Ashgill, but most Ordovician genera persist into the Silurian. However, very few Ordovician species range into the Llandovery.

Diagnostic Llandovery chitinozoans

Silurian chitinozoans (Figs 13–27) are more widely distributed than Ordovician ones (Laufeld 1979). Endemic chitinozoan faunas do occur, but not to the same extent as during the Ordovician. However, there is a difference in chitinozoan assemblages between north Africa, Anticosti Island and Baltoscandia. Chitinozoans from the type Llandovery are poorly preserved and the diversity seems to be low (K. Dorning, personal communication 1985). On the other hand, Telychian faunas from Great Britain show a striking similarity to contemporaneous faunas in Baltoscandia (Aldridge *et al.* 1979; Dorning 1981; Mabillard & Aldridge 1985).

Figs 13–27 Selected Llandovery Chitinozoa. 13, *Coronochitina fragilis*, Juuru Stage (early Llandovery), Ohesaare boring (466·5 m), Estonia; SEM × 300. 14, *Conochitina armillata*, middle-late Llandovery, boring D1-31 (1895–1896 m), Libya; SEM × 160. 15, *Conochitina edjelensis elongata*, middle-late Llandovery, boring E1-81 (606–612 m), Libya; SEM × 160. 16, *Eisenackitina dolioliformis*, Restevo Beds (late Llandovery), Podolia, U.S.S.R.; SEM × 420. 17, *Conochitina aspera*, Juuru Stage (early Llandovery), Ikla boring (514·6 m), Estonia; SEM × 430. 18, *Conochitina proboscifera*, Upper Visby Beds (early Wenlock), Gotland, Sweden; SEM × 70. 19, *Pterochitina dechaii*, middle-late Llandovery, boring D1-31 (1895–1896 m), Libya; SEM × 325. 20, *Plectochitina pseudoagglutinans*, middle-late Llandovery, boring A1-81 (1154–1161 m), Libya; SEM × 195. 21, *Conochitina electa*, Raikküla Stage (middle Llandovery), Emmaste boring (41·2 m), Estonia; SEM × 160. 22, *Angochitina longicollis*, Lower Visby Beds (late Llandovery), Gotland, Sweden; SEM × 175. 23, *Conochitina iklaensis*, Raikküla Stage (middle Llandovery), Ikla boring (492·0 m), Estonia; SEM × 160. 24, *Coronochitina maennili*, Raikküla Stage (middle Llandovery), Ikla boring (462·9 m), Estonia; SEM × 160. 25, *Ancyrochitina convexa*, Raikküla Stage (middle Llandovery), Ruhnu boring (536·0 m), Estonia; SEM × 300. 26, *Desmochitina densa*, Upper Visby Beds (early Wenlock), Gotland, Sweden; SEM × 345. 27, *Ancyrochitina laevensis*, Juuru Stage (early Llandovery), Laeva boring (122·5 m), Estonia; SEM × 300.

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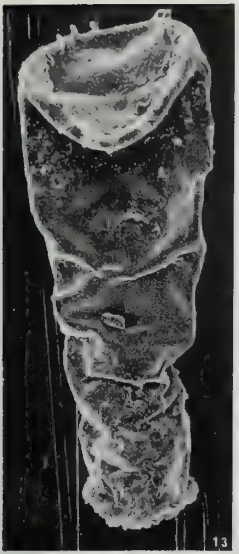
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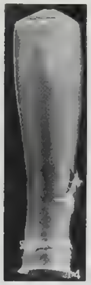
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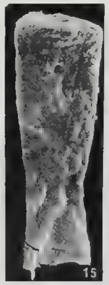
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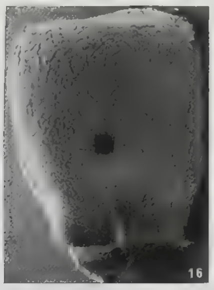
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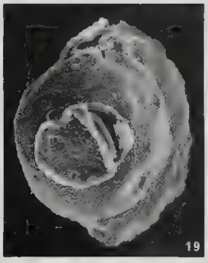
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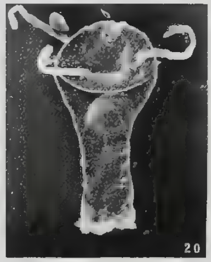
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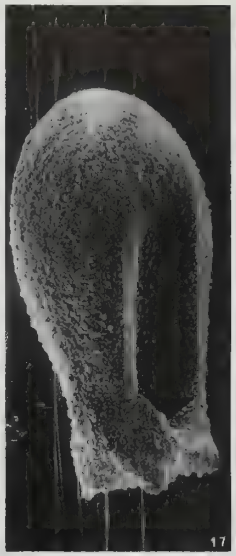
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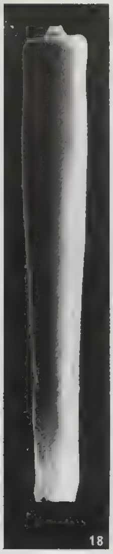
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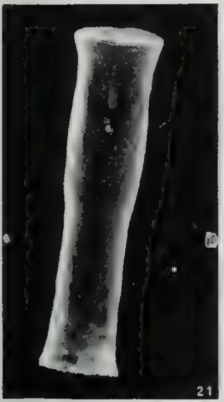
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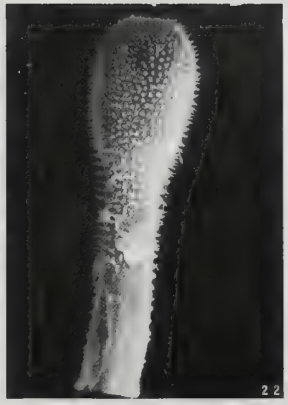
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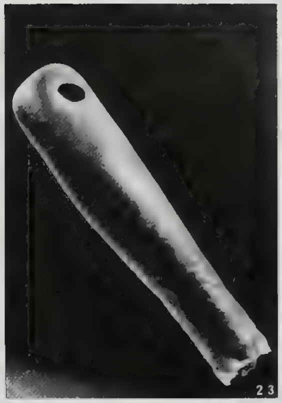
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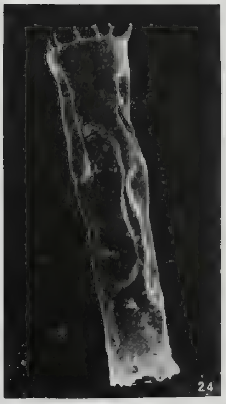
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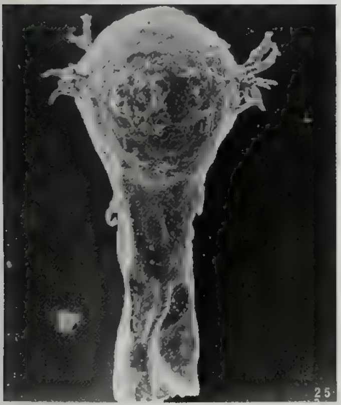
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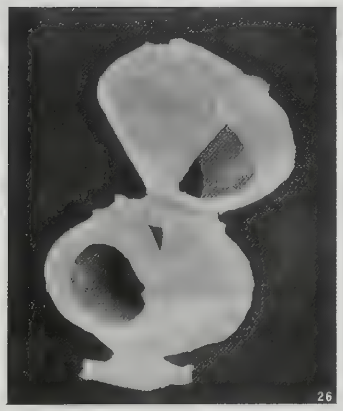
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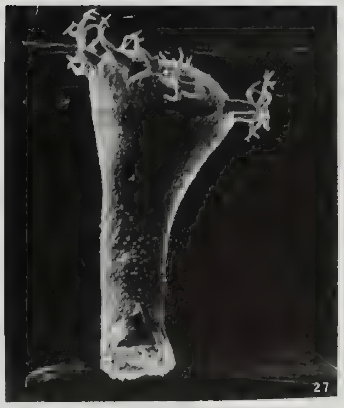
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The appearance of *Ancyrochitina laevensis* and *Coronochitina fragilis* indicates lowermost Rhuddanian strata (Nestor 1980*a*). Otherwise pre-*cyphus* beds have a low chitinozoan diversity, and, apart from *Conochitina aspera*, there are very few diagnostic species above the *acuminatus* Zone (Fig. 1). In the *cyphus* Zone *Conochitina iklaensis* occurs, and is joined in the topmost part by *Coronochitina maennili* (Nestor 1980*a*). These two species disappear in the *sedgwickii* Zone (Nestor 1980*a*; own observations) together with *Conochitina edjelensis*, a useful representative of the Aeronian. The lowermost Aeronian is characterized by the presence of *Ancyrochitina convexa* (Nestor 1980*b*). *Eisenackitina dolioliformis* and *Conochitina emmastiensis* (Nestor 1982*a*) have their first appearance in the *sedgwickii* Zone and range into the Wenlock. A very characteristic chitinozoan assemblage occurs in the *griestoniensis* Zone, consisting of *Angochitina longicollis*, *Conochitina proboscifera* and *Desmochitina densa* and is widely distributed (Dorning 1981; Nestor 1982*b*; Verniers 1982; Mabillard & Aldridge 1985; etc.).

The presence of Baltoscandian species among north African chitinozoan assemblages makes it possible to determine the stratigraphical ranges of some north African taxa (Paris, in press), such as *Plectochitina pseudoagglutinans* and *Conochitina vitrea* (Hill *et al.* 1985). These species range from the lower Rhuddanian to the upper Aeronian (Fig. 1). Two other species, *Conochitina armillata* and *Pterochitina deichaii*, range from the mid-Aeronian to the mid-Telychian (Paris, in press).

Remarks on the boundary chitinozoans

In general the abundance and diversity of chitinozoans are comparatively low at the Ordovician–Silurian boundary, irrespective of geographic area. This is probably due to the Gondwana glaciation, which led to a eustatic sea-level drop, and the subsequent deposition of shallow-water sediments in many cratonic successions. Chitinozoans are usually rare or absent in rocks deposited in very shallow water (Laufeld 1974; Grahn & Bergström 1984, 1985). If chitinozoans are present in these rocks, planktic forms often dominate and these were probably transported inshore by currents and waves. This is demonstrated in the Belfast Beds of early Llandovery age in the Cincinnati Region, where the planktic genus *Ancyrochitina* constitutes about 99% of the chitinozoan fauna (Grahn & Bergström 1985).

Acknowledgements

I am indebted to Sven Laufeld (Uppsala) and Florentin Paris (Rennes) for critical reading and improvements of the manuscript. Aicha Achab (Ste-Foy), Viiu Nestor (Tallinn), Florentin Paris (Rennes), Sven Laufeld (Uppsala), Jaak Nõlvak (Tallinn), and Alain Le Herissé (Brest) provided me with SEM-pictures, and Francine Martin (Bruxelles) and Merrell A. Miller (Tulsa) with samples. Karin Feltzin (Stockholm) finished my line drawing and Richard J. Aldridge (Nottingham) checked the English. My sincere thanks to all these friends.

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Conodont biostratigraphy of the Uppermost Ordovician and Lowermost Silurian

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Synopsis

A review of the conodont biostratigraphy of the Ordovician–Silurian boundary sections in North America, Europe, and Asia shows that virtually all sections are either incomplete stratigraphically or have intervals from which no diagnostic conodonts are known. The best known conodont succession across the systemic boundary is on Anticosti Island, where, however, the precise level of the boundary remains unknown because of the absence of diagnostic graptolites. Ordovician and Silurian conodont faunas differ greatly and there is conclusive evidence that a conspicuous turnover in the conodont faunas took place globally in the systemic boundary interval. This turnover involved the replacement of a fauna of Ordovician aspect containing more than 25 genera with one of Silurian aspect having fewer than 15 genera, eight of which are known also from the Ordovician. A few coniform conodont species survived this extinction event, but we have identified only one species with compound elements in the apparatus that may range from the uppermost Ordovician to lowermost Silurian; however, even in the case of this form, there is some question whether we are dealing with the same species in both systems. The dating of the conodont faunal turnover in terms of standard graptolite zones is still somewhat uncertain, but available data suggest that it occurs in an interval in the upper *G. persculptus* Zone but below the systemic boundary. This extinction event is probably a result of the Saharan glaciation. In those cases where the origin of the Llandovery stocks is known or can be postulated, they appear to be derived, in almost all cases, from stocks that inhabited the tropical waters of the Midcontinent Province during the Ordovician. It is concluded that further studies are urgently needed, particularly to date exactly the conodont faunal turnover and to define the Ordovician–Silurian boundary in terms of the conodont succession.

Introduction

Extensive research during the last few decades has firmly established conodonts as a key zone fossil group in Ordovician and Silurian rocks. The conodont zone successions now in use within each of these systems provide a stratigraphical resolution which in many cases is superior to that of other fossil groups, also including the graptolites. Furthermore, the fact that conodonts are present in rocks representing the whole range of marine depositional environments from basinal to intertidal, or even supratidal, makes them very useful for both local and regional biostratigraphical work. This is particularly the case in the shallow-water carbonate deposits that occupy vast areas on the cratons of all continents except Africa and Antarctica but which contain only few and scattered occurrences of zonal graptolites.

In view of the significance of conodonts as zonal fossils in Ordovician–Silurian strata, it is hardly surprising that they played a major role in the lengthy discussions about the Ordovician–Silurian boundary which were carried out within the Ordovician–Silurian Boundary Working Group of the I.U.G.S Commission on Stratigraphy. Although it was ultimately decided to define this systemic boundary on graptolites, the absence of diagnostic graptolites in many boundary sections, particularly the cratonic ones, makes it necessary to use other fossils for establishing the precise level of the systemic boundary. Conodonts have great potential to serve in this capacity. The purpose of the present contribution is to summarize and assess currently available conodont evidence that has bearing on the recognition and definition of the Ordovician–Silurian boundary. Although we attempt global coverage, we will concentrate on North America and Europe, where the most detailed studies have been carried out and from which we have not only easily accessible information but also personal field experience of most of the important boundary sections.

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Upper Ordovician–Lower Silurian Conodont Zonations

The striking faunal provincialism of Late Ordovician conodonts (Barnes *et al.* 1973; Bergström 1973; Sweet & Bergström 1974, 1984; Dzik 1983) has necessitated the use of separate biostratigraphical zonal schemes for the North Atlantic and Midcontinent provinces. Although Sweet & Bergström (1984) recently introduced more refined provincial units for the Upper Ordovician of North America and Europe, in the present contribution, which is global in scope, we use only these two provinces. Provincialism was not conspicuous during the Early Silurian but several slightly different zonal schemes have been proposed. However, eventually it may be possible to use a single zonal scheme globally for this part of the succession.

The Middle and Upper Ordovician zone succession for the North Atlantic Province developed by Bergström (1971*a*, 1971*b*, 1978, 1983, 1986) has been tested and used by many other authors, e.g. Dzik (1976, 1983), Harris *et al.* (1979), Orchard (1980), and Schönlaub (1971, 1980). This zonal scheme (Fig. 1) is based on the evolutionary lineage of *Amorphognathus.* The successive zones of *A. tvaerensis*, *A. superbus*, and *A. ordovicicus* covers the Caradoc–Ashgill interval. The *A. tvaerensis* Zone has three named subzones but no attempt has yet been made to subzone the *A. superbus* and *A. ordovicicus* Zones although the restricted stratigraphical range of some taxa (e.g. *A. complicatus*, *Hamarodus europaeus*, *Sagittodontina robusta*; Bergström 1983: fig. 1) may eventually allow this (cf. Orchard 1980).

Conodont biostratigraphical classification of the Upper Ordovician of the Midcontinent Province was first developed as a sequence of faunas characteristic of particular stratigraphical intervals (Sweet *et al.* 1971; Sweet & Bergström 1976; McCracken & Barnes 1981). The interval of Faunas 10–13 covered the Cincinnatian Series. Later work by Sweet (1979*a*, 1979*b*, 1984) using graphic correlation methods has led to the establishment of a Composite Standard Section and a formal zonal scheme with the successive *Belodina confluens*, *Oulodus velicuspis*, *O. robustus*, *Aphelognathus grandis*, *A. divergens*, and *A. shatzeri* Zones. Because of regional migration of North Atlantic Province faunal elements into the Midcontinent Province during the Late Ordovician (Sweet *et al.* 1971: fig. 3), it is possible to tie some of the zonal boundaries of these two provincial zone schemes (Sweet 1984: fig. 2). Other studies documenting and supporting this scheme include those of Nowlan & Barnes (1981), McCracken & Barnes (1981, 1982) and Nowlan *et al.* (in press). Outside North America, studies of cratonic conodont faunas have been undertaken by, among others, Moskalenko (1983), An (1981), and An *et al.* (1983), and a formal zonation has been proposed for Siberia (Moskalenko 1983). It is possible that other low latitude Ordovician plates (e.g. Kazakhstan, north China and Australia) may require separate zonal schemes because their conodont faunas include many endemic elements.

The first attempt to develop a conodont zonal scheme for the Lower Silurian was by Walliser (1964, 1971) from work in the Carnic Alps. Work in this area was later undertaken by Schönlaub (1971, 1980). Following descriptions of faunas from other regions, it gradually became apparent that the Carnic Alps standard sequences were stratigraphically incomplete. Aldridge (1972, 1975) established a new zonation in the Welsh Borderland, but non-productive clastics in the lowermost Silurian there prevented the establishment of a complete zonal succession through the Llandovery. In North America, Barrick (1977), Barrick & Klapper (1976), Cooper (1975, 1980), Fåhraeus & Barnes (1981), Helfrich (1980), LeFèvre *et al.* (1976), McCracken & Barnes (1981), Nowlan (1983), Pollock *et al.* (1970), Rexroad (1967), Nicoll & Rexroad (1971), and Uyeno & Barnes (1983), among others, have documented faunas from important sequences. Elsewhere, studies of Early Silurian conodonts include those of Männik (1983) in Severnaya Zemlya, USSR, Lin (1983) in China, and Igo & Koike (1968) from Malaysia.

As a result of these studies, two Lower Silurian conodont zone schemes have evolved for North America and Europe (Fig. 1) and another for China. However, the phylogenies of important lineages, such as those of *Icriodella*, *Distomodus* and *Oulodus*, have yet to be fully documented, and the precise ranges of several key species, including platform taxa, are not yet established. Once these have been clarified, particularly in sequences such as those on Anticosti Island, a single zonal scheme should be applicable to most areas. There is also an urgent need for further documentation of the conodont species succession across the Ordovician–Silurian

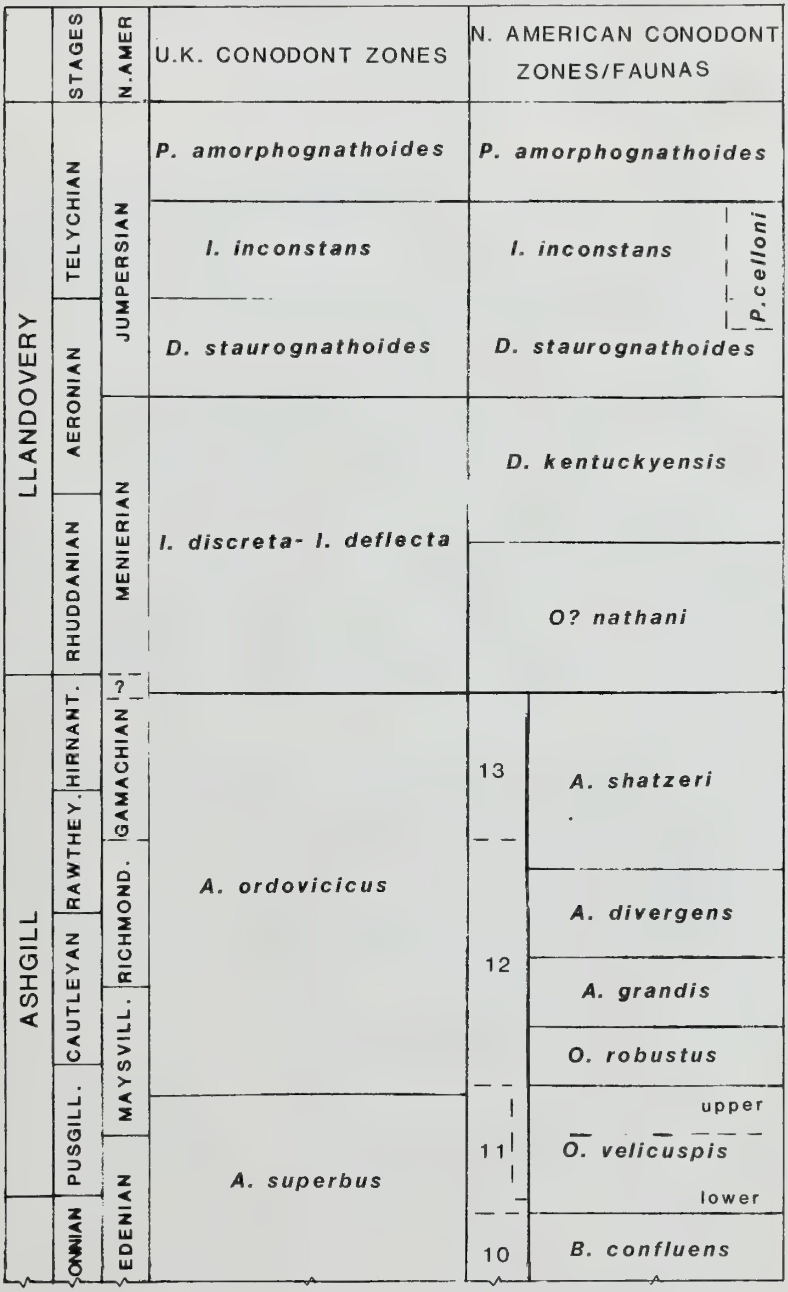
STAGES
N.AMER
U.K. CONODONT ZONES
N. AMERICAN CONODONT ZONES/FAUNAS
LLANDOVERY
ASHGILL
TELYCHIAN
AERONIAN
RHUDDANIAN
HIRNANT.
RAWTHEY.
CAUTLEYAN
PUSGILL.
ONNIAN
JUMPERSIAN
MENIERIAN
?
GAMACHIAN
RICHMOND.
MAYSVILL.
EDENIAN
P. amorphognathoides
I. inconstans
D. staurognathoides
I. discreta- I. deflecta
A. ordovicicus
A. superbus
P. amorphognathoides
I. inconstans
P. celloni
D. staurognathoides
D. kentuckyensis
O? nathani
13
A. shatzeri
12
A. divergens
A. grandis
O. robustus
11
upper
O. velicuspis
lower
10
B. confluens

Fig. 1 Late Ordovician–Early Silurian chronostratigraphy and conodont zonation for U.K. and North America. For new terminology of Silurian chronostratigraphy see Barnes (in press) and Holland (1985); for zones see references in text.

boundary. Even though faunal provincialism is much reduced in the Lower Silurian when compared to the Upper Ordovician, the conodont faunas exhibit considerable differentiation horizontally; hence there are significant biofacies differences between nearshore and basinal environments (e.g. Aldridge & Mabillard 1981), and community patterns across shelf environments can be deduced (e.g. LeFèvre *et al.* 1976; McCracken & Barnes 1981; Nowlan 1983; Uyeno & Barnes 1983).

North America

Conodont studies of strata close to, or across, the boundary interval have been undertaken in many regions in North America, including Anticosti Island, Gaspé, the Michigan, Hudson Bay, Williston and Illinois basins and adjoining arch areas, the western Midcontinent, the Cordillera, Arkansas–Oklahoma, and the Canadian Arctic and its extension into northern Greenland. The best section currently known is on Anticosti Island, Québec, where there is a continuous and continuously fossiliferous sequence across the systemic boundary. Elsewhere, there is a stratigraphical hiatus in the boundary interval, or the faunal sequence is incomplete.

The Anticosti Island conodont sequence (Fig. 2) has been documented by Nowlan & Barnes (1981), McCracken & Barnes (1981), Fåhraeus & Barnes (1981), Uyeno & Barnes (1983), and Barnes (this volume). Conodont Fauna 13 is developed in Gamachian strata, and the *Oulodus*? *nathani*, *Distomodus kentuckyensis*, *D. staurognathoides*, *Icriodella inconstans*, and *Pterospathodus amorphognathoides* Zones (Fig. 1) are recognized in Llandovery strata. These studies are based on intensive sampling and on the investigation of nearly 100 000 superbly preserved conodonts. Conodont Fauna 13 of McCracken & Barnes (1981), which contains the distinctive genus *Gamachignathus*, is associated with Ordovician macrofossils such as *Vellamo* and aulacerids. Through the overlying *O.*? *nathani* Zone there is a sequential occurrence of Silurian brachiopods (*Zygospiraella*, *Stricklandia*, *Virgiana*) and the trilobite *Acernaspis* (Lespérance 1985). From one locality on eastern Anticosti Island Cocks & Copper (1981) reported a *Hirnantia* brachiopod fauna just below a level where Nowlan (1982) recovered conodonts of Silurian aspect.

On the Gaspé Peninsula (Fig. 2), Québec, the White Head Formation exhibits a faunal sequence similar to that of Anticosti Island. *Gamachignathus* (Fauna 13) is known from Unit 4 of this formation, the *Hirnantia* fauna and the *Mucronaspis* fauna are well developed and associated with *G. persculptus* Zone graptolites in Unit 5, and *Acernaspis* occurs with Silurian conodonts (*D. kentuckyensis*) in Unit 6 (Nowlan 1981, 1983; Lespérance 1985). In another part of Gaspé, the *O.*? *nathani* Zone has been recognized in the Clemville Formation (Nowlan 1983).

On Anticosti Island there is a marked faunal change with a rapid replacement of a diverse Ordovician conodont fauna with a distinctive, but less diverse, Silurian fauna. In an interval up to two metres thick, a few Ordovician taxa co-occur with species of Silurian aspect. Unfortunately, the absence of graptolites diagnostic of the *P.*? *acuminatus* Zone in the Anticosti Island succession makes it impossible to establish the precise level of the systemic boundary, and the relations between the faunal turnover and this level. The fact that the uppermost interval of Fauna 13 has a *Hirnantia* fauna and graptolites of the *G. persculptus* Zone on Gaspé (Lespérance 1985) shows that the conodont fauna below the turnover interval is of pre-Silurian age, and the systemic boundary must be at a higher stratigraphical level in Anticosti. Lespérance (1985) suggested that the appearance of *Acernaspis* may be coeval with the base of the *P.*? *acuminatus* Zone and hence mark the systemic boundary; however, as noted below, the reliability of the appearance of this genus regionally as a guide to the boundary level needs confirmation, and its appearance on Anticosti Island might be at a higher stratigraphical level than in some other areas.

In Ontario and Michigan in the Great Lakes region, conodont studies have revealed the existence of a hiatus at the boundary that spans the Gamachian Stage and possibly parts of the Richmondian and early Llandovery as well (cf. Barnes & Bolton, this volume). Fauna 13 and the *O.*? *nathani* Zone are not recognized in this area. A similar hiatus exists to the north in the Hudson Bay Basin (LeFèvre *et al.* 1976) and to the south in the Cincinnati Region (cf. Sweet

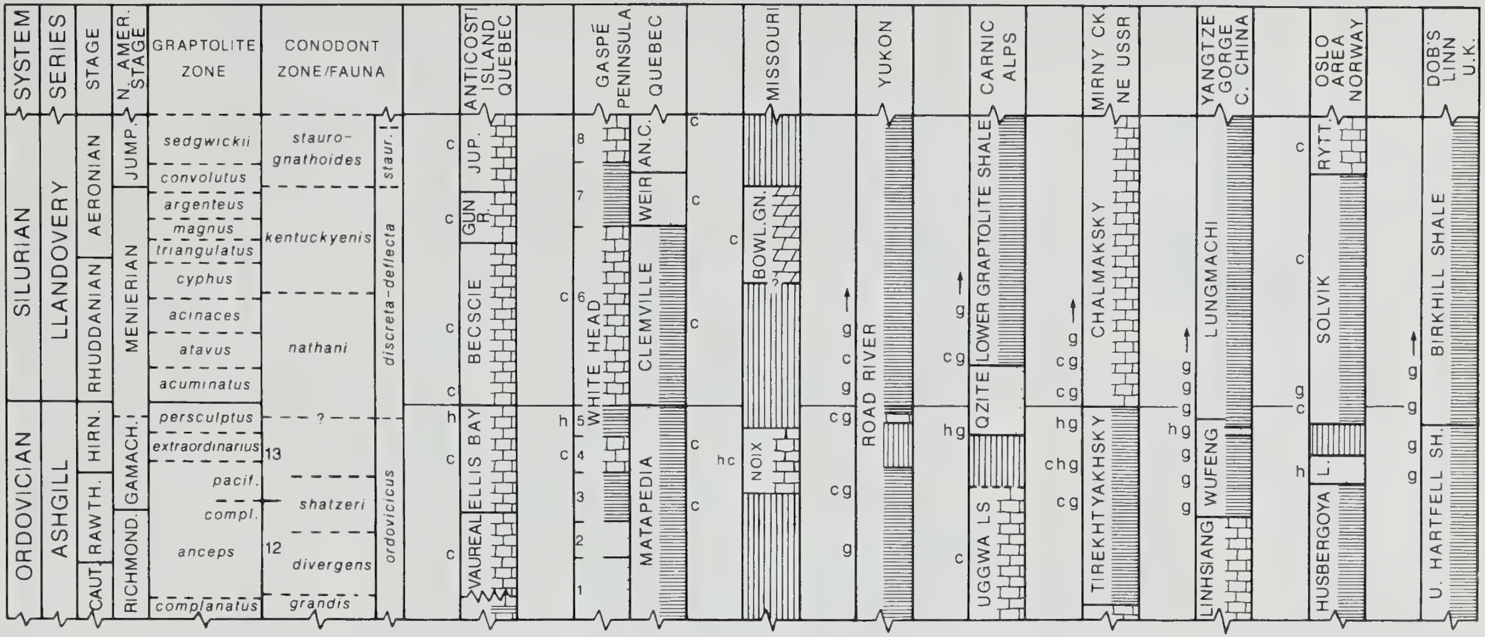
SYSTEM
SERIES
STAGE
N. AMER. STAGE
GRAPTOLITE ZONE
CONODONT ZONE/FAUNA
ANTICOSTI ISLAND QUEBEC
GASPÉ PENINSULA
QUEBEC
MISSOURI
YUKON
CARNIC ALPS
MIRNY CK NE USSR
YANGTZE GORGE C. CHINA
OSLO AREA NORWAY
DOB'S LINN U.K.
SILURIAN
ORDOVICIAN
LLANDOVERY
ASHGILL
AERONIAN
RHUDDANIAN
HIRN.
RAWTH.
CAUT.
JUMP.
MENIERIAN
GAMACH.
RICHMOND.
sedgwickii
convolutus
argenteus
magnus
triangulatus
cyphus
acinaces
atavus
acuminatus
persculptus
extraordinarius
pacif.
compl.
anceps
complanatus
stauro-gnathoides
kentuckyenis
nathani
?
13
shatzeri
12
divergens
grandis
staur.
discreta–deflecta
ordovicicus
JUP.
GUN R.
BECSCIE
ELLIS BAY
VAUREAL
WHITE HEAD
1
2
3
4
5
6
7
8
AN.C.
WEIR
CLEMVILLE
MATAPEDIA
BOWL.GN.
NOIX
ROAD RIVER
LOWER GRAPTOLITE SHALE
QZITE
UGGWA LS
CHALMAKSKY
TIREKHTYAKHSKY
LUNGMACHI
WUFENG
LINHSIANG
RYTT.
SOLVIK
L.
HUSBERGOYA
BIRKHILL SHALE
U. HARTFELL SH.
c
g
h
cg
hg
chg
hc

Fig. 2 Key stratigraphical sections across the Ordovician–Silurian boundary showing dominant lithologies of formation and their biostratigraphical correlation based on presence of conodont (c), graptolite (g), and *Hirnantia* (h) faunas.

1979*a*, 1984; Grahn & Bergström 1985). The faunas of the latter region have been well documented in the last decades by W. C. Sweet and co-workers for the Ordovician and C. B. Rexroad and co-workers for the Llandovery. Further to the west, in the Williston Basin, no Gamachian Fauna 13 has been recognized (Sweet 1979*b*; Barnes, unpublished collections from Manitoba), and the earliest Silurian conodonts compare well with those from the Manitoulin Formation of Ontario discussed by Barnes & Bolton (this volume). These conodont successions from the Midcontinent Region suggest that a major regression left the North American craton largely emerged for at least the duration of the Gamachian, and possibly longer, at least in some areas. The only exceptions to this, that is, areas where youngest Ordovician conodonts are present, are in marginal basins (e.g. Anticosti Island), some intracratonic troughs (e.g. in Arkansas–Missouri–Oklahoma; see Bergström & Boucot, this volume), outer miogeoclinal areas (e.g. Utah and Nevada), and regions having offshore basin and slope deposits (e.g. Gaspé and Arctic Canada).

In some Midcontinent areas (Fig. 2), incomplete stratigraphical successions produce intriguing conodont faunas of latest Ordovician age. Such faunas are known from the Cason Oolite of Arkansas (Craig 1969, 1986; Barrick 1986), the Noix Oolite and Girardeau Limestone of Missouri (Satterfield 1975; McCracken & Barnes 1982), and the Keel Formation of Oklahoma (Barrick 1986). These units yield sparse faunas characterized by *Noixodontus girardeauensis* (Satterfield). McCracken & Barnes (1982) assigned a Fauna 12 (Richmondian) age to the Noix fauna, but Barrick (1986) suggests that the presence of a *Hirnantia* fauna in several of these units indicates a latest Ordovician (late Gamachian, Hirnantian, Fauna 13) age. In the Yukon (Fig. 2), Lenz & McCracken (1982) recorded both *Noixodontus* and *Gamachignathus* in strata referred to the *Pacificograptus pacificus* Zone (the upper *Climacograptus supernus* Zone, equivalent to the lower part of the interval of the *Hirnantia* fauna in China; Lenz & McCracken 1982: fig. 6). In the Yukon, the overlying *Climacograptus extraordinarius* Zone is not recognized and that interval may be represented by a hiatus. The latest Ordovician *Glyptograptus persculptus* Zone is identified only with question, but significantly a Silurian conodont fauna is recorded from 6·3–13·3 m below the top of the *G. persculptus* Zone? in the Pat Lake section (Lenz & McCracken 1982, Appendix). With a hiatus below the *G. persculptus* Zone?, it is possible that only the uppermost part of that zone is present in the succession.

In the Canadian and Greenland Arctic regions, several conodont studies have been completed, or are under way, but little has been published to date. Preliminary results (Mayr *et al.* 1980) suggest the presence of a regionally developed hiatus in the systemic boundary interval. This is certainly the case in the carbonate platform facies (e.g., the Allen Bay Formation) and probably in the basinal facies as well, where the *G. persculptus* Zone has not been recognized.

Finally, Leatham (1985) has described a section in carbonate facies across the systemic boundary interval in the Great Basin. Absence of graptolites precludes recognition of the precise level of the systemic boundary. However, Leathan recognized an interval with mixed faunas between typical Ordovician and typical Silurian faunas, but he was inclined to believe that these mixed faunas were due to stratigraphical leaks or reworking of Ordovician conodonts into basal Silurian strata near an unconformity associated with the systemic boundary. In central Nevada, Ross *et al.* (1979) interpreted the Hanson Creek Formation as ranging without significant gap from the Late Ordovician to the Early Silurian. Fauna 13 seems to be represented in their collections but because they do not describe their Silurian conodonts, it is not clear how the conodont faunal succession is developed in the boundary interval.

Great Britain

No continuous section across the Ordovician–Silurian boundary developed in a facies suitable for conodont extraction is known from the British Isles. The boundary stratotype at Dob's Linn, Scotland (Fig. 2), as well as the lowermost part of the Llandovery reference standard in south Wales, are both unpromising for conodont work. A few conodonts have been recovered from shale bedding planes at the boundary stratotype, Dob's Linn (Barnes & Williams, this volume), and a single conodont collection is known from the lowermost Llandovery of the type area (Cocks *et al.* 1984). Efforts to collect from strata near the systemic boundary elsewhere in Britain have not been very successful; hence, only two productive samples are known from the Hirnantian (Bergström & Orchard 1985), none of them with very diagnostic species although the faunas are clearly of Ordovician aspect. Apparently, as in Scandinavia, the Hirnantian rocks in Britain are very poor in conodonts.

Currently available information about British early Llandovery conodonts derives largely from the work by Aldridge and co-workers. As noted by Aldridge (1985), very few conodonts are currently known from the Rhuddanian although a sample from the lower part of the stage at Llandovery contained a species association diagnostic of Aldridge's (1972) *Icriodella discreta–I. deflecta* Zone (Cocks *et al.* 1984). Aeronian strata in Wales and the Welsh Borderland have yielded taxonomically varied species associations (Aldridge 1985), which include *Kockelella*? *abrupta*, *Ozarkodina oldhamensis*, *O. hassi*, and *Pterospathodus*? *tenuis*. The upper Aeronian is characterized by the appearance of *Distomodus staurognathoides*, *Oulodus*? *fluegeli*, *Pseudooneotodus tricornis*, and *Kockelella ranuliformis*.The interval having this species association is referable to the *Distomodus staurognathoides* Zone (Aldridge 1972).

Scandinavia

The few sections in Sweden (Västergötland, Scania) and Denmark (Bornholm) where the base of the *Parakidograptus*? *acuminatus* Zone, and hence the base of the Silurian, can be recognized are all in dark shale facies from which no conodonts have been recovered. In other sections, shallow-water strata with the *Hirnantia* fauna (Bergström 1968) are overlain, in places unconformably, by Llandovery age shales and mudstones. In Sweden, the Ashgill conodont faunas are known from several sections (Bergström 1971*a*; Sweet & Bergström 1984) but the early Llandovery ones are virtually unknown. No conodonts have been recorded from the systemic boundary interval in Denmark.

Biostratigraphically well controlled lower Llandovery successions have recently been described from the Oslo region, Norway (Fig. 2). The conodont succession there is particularly significant because it can be tied to the distribution patterns of key graptolites and shelly fossils (Aldridge & Mohamed 1982). As is the case in Sweden, rocks of latest Ordovician (Hirnantian) age have produced very few conodonts, the only reasonably common species being a form close to, if not identical with, *Ozarkodina oldhamensis*, which is also characteristic of coeval strata in Sweden (Bergström 1971*a*: fig. 4:11). Absence of close graptolite control makes it impossible to establish the precise level of the systemic boundary in the Oslo region, but the graptolites indicative of the upper *Glyptograptus persculptus* Zone or lower *P.*? *acuminatus* Zone present in the lower Solvik Formation (Howe 1982) suggest that the systemic boundary is close to the base of that unit, which is separated from the underlying Hirnantian strata by what appears to be a minor gap. The recent suggestion that the appearance of the trilobite *Acernaspis* is coeval with the base of the *P.*? *acuminatus* Zone is not well supported by the conditions in the Oslo region where this genus makes it appearance in the middle Solvik Formation (6bα) in an interval that on graptolite evidence appears to be no older than the *Monograptus atavus* Zone (Howe 1982).

A summary of the conodont, shelly fossil, and graptolite biostratigraphy of the lower Llandovery of the Oslo region is given in Fig. 3. The faunal succession is quite similar to that of the Anticosti Island (Barnes & McCracken 1981; Lespérance 1985), Gaspé (Nowlan 1983; Lespérance 1985), and the Rhuddanian and lower Aeronian of Britain (Aldridge 1985; Cocks *et al.* 1984). In the lowermost Llandovery of the Oslo region, the presence of *Oulodus*? cf. *O.*? *nathani* strongly suggests that the *Oulodus*? *nathani* Zone can be recognized (Aldridge & Mohamed 1982), which is overlain by the *Distomodus kentuckyensis* Zone. In the uppermost part of the Solvik Formation, representatives of *Distomodus staurognathoides* and other species of the *D. staurognathoides* Zone make their entrance, which suggests correlation with the middle Aeronian of Britain (Aldridge 1975) and the lower part of the Jupiter Formation of Anticosti Island (Uyeno & Barnes 1983). Although the Llandovery conodont succession of the Oslo region is one of the best biostratigraphically controlled in the world, it unfortunately

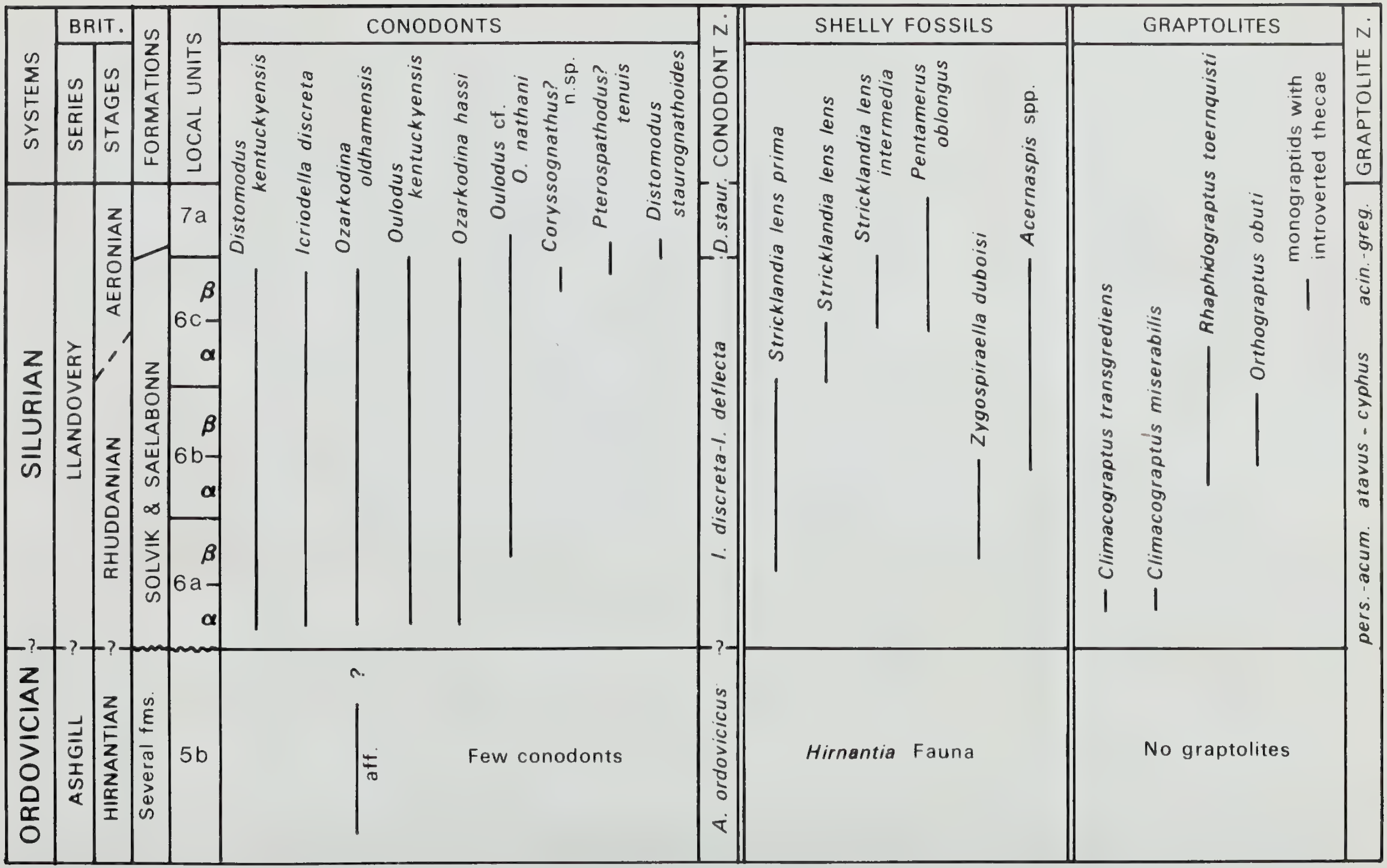

Fig. 3 Comparison of stratigraphical ranges of key conodonts, shelly fossils, graptolites, and conodont and graptolite zones in the lower Llandovery of the Oslo region, Norway. Based on many sources, particularly Howe (1982) and Aldridge & Mohamed (1982). Note that the *Hirnantia* fauna-bearing upper Ashgill (5b), which is separated from overlying rocks by a minor unconformity, has yielded only a few conodonts and no diagnostic graptolites.

provides little information about the conodont sequence right across the systemic boundary interval, despite the fact that much of this interval is developed in calcareous rocks that are readily digestible in weak acids.

Carnic Alps and nearby areas in Austria and Italy

The Cellon section in the Carnic Alps has become classic as the reference standard of much of the Silurian conodont zone succession (Walliser 1964) but this border region between Austria and Italy has several other important sections that include Late Ordovician as well as Early Silurian strata (Fig. 2) (Schönlaub 1969, 1971, 1979, 1980; Jaeger & Schönlaub 1977; Jaeger *et al.* 1975; Serpagli 1967; Vai 1971; Flajs & Schönlaub 1976). Because graptolites diagnostic of the *P.*? *acuminatus* Zone are unknown in the Carnic Alps, the precise level of the base of the Silurian cannot be determined in sections with more or less continuous sequence. In other sections, Silurian or younger strata rest unconformably on Ordovician beds and the systemic boundary coincides with a conspicuous stratigraphical gap.

Many of the conodont data available from this region pertaining to the Ordovician–Silurian boundary interval consist of lists of species, but there are also published descriptions and illustrations of Ashgill (Serpagli 1967; Flajs & Schönlaub 1976) and Llandovery (Walliser 1964; Schönlaub 1971) conodonts. Sweet & Bergström (1984) suggested some updating of the taxonomy of Ashgill species and additional taxonomic work on some of the faunas is clearly needed.

The most distinctive Ashgill age unit in the Carnic Alps is an argillaceous limestone a few metres thick, the Uggwa (Uqua) Limestone (= Tonflaserkalk). Although its conodont fauna, which was monographed by Serpagli (1967), includes some species currently unknown outside Austria and Italy, it is clearly of Ordovician rather than Silurian aspect and represents the *Amorphognathus ordovicicus* Zone. Some of its characteristic genera include *Amorphognathus*, *Ansella*, *Birksfeldia*, *Drepanoistodus*, *Hamarodus*, *Plectodina*, *Protopanderodus* and *Scabbardella*, which are all restricted to the Ordovician. In several sections, the Uggwa Limestone is followed by a prominent stratigraphical gap that may represent a portion of the Silurian (or more) and possibly also the uppermost Ordovician. At other sections, a part of this gap is filled by calcareous sandstones and dark shales, commonly referred to as the 'Untere Schichten', that locally, for instance at the Cellon section, contain megafossils of the *Hirnantia* fauna associated with Ashgill conodonts. Walliser (1964) classified the 'Untere Schichten' as the upper part of his Bereich 1 and referred this unit to the Lower Silurian. We believe that most, if not all, of the 'Untere Schichten' belongs to the uppermost Ordovician, if one follows the practice of having the systemic boundary at the base of the *P.*? *acuminatus* Zone.

As shown by Walliser (1964), the beds on the top of the 'Untere Schichten' at Cellon contain conodonts (*Apsidognathus tuberculatus*, *Distomodus staurognathoides* and *Pterospathodus celloni*) of the *P. celloni* Zone, and a similar fauna is known also from beds just above the Ashgill age limestone at the Mount Seewarte section (Schönlaub 1971, 1980). At both these sections, the stratigraphical hiatus associated with the systemic boundary includes two-thirds of the Llandovery (Rhuddanian and Aeronian stages). On the other hand, at other localities, such as the Feistritzgraben section (Jaeger *et al.* 1975; Schönlaub 1980), the Uggwa Limestone is directly overlain by dark shales that contain *Glyptograptus* cf. *G. persculptus* near their base. This suggests a much smaller, if any, stratigraphical gap above the limestone, and the systemic boundary is evidently at an unknown level in the clastic succession above the graptolite-bearing interval.

Although earliest Silurian, and perhaps also latest Ordovician, conodonts are unknown from the sections in the Carnic Alps and nearby regions, this area is of interest in discussions about the conodont biostratigraphy near the systemic boundary because of its rich Ashgill and middle and late Llandovery conodont faunas. Furthermore, in view of the local variations in both lithological and stratigraphical development near the systemic boundary, it is not excluded that further studies may lead to the discovery of stratigraphically more complete sections in a lithology suitable for extraction of conodonts than those now known.

Other areas

Outside North American and Europe, latest Ordovician and/or earliest Silurian conodonts are known from Siberia, China and Malaysia. In her review of the Ashgill conodont biostratigraphy of the Siberian Platform, Moskalenko (1983) recognized an *Aphelognathus pyramidalis* Zone in the topmost part (the Burian Stage) of the Ordovician but she noted that the succession is terminated by an erosional unconformity. Apart from the zonal index, the low-diversity and apparently largely endemic conodont fauna includes, among others, *Acanthodina nobilis*, *A. variabilis*, and *Acanthodus compositus* (Moskalenko 1973). Männik (1983) recorded a conodont succession through the Silurian of Severnaya Zemlya. The lowermost unit, the Vodopad Formation, yielded in its lower part *Ozarkodina oldhamensis*, *Icriodella* cf. *I. deflecta*, and *Oulodus*? cf. *O. kentuckyensis*, among others. This interval was referred to the *I. discreta–I. deflecta* Zone and interpreted to be of late Rhuddanian to early Aeronian (= Idwian in Männik) age. The similarity to coeval faunas in the Oslo region and eastern Canada is striking.

In China, the uppermost Ordovician, where present, is in most places developed in a lithology unsuitable for conodont extraction, and it has yielded only a few undiagnostic species (An 1981). Shelly facies of Llandovery age produce taxonomically varied and well preserved conodonts such as those from the Guizhou Province recorded by Zhou *et al.* (1981; also cf. Lin 1983) that provide correlation with the early Llandovery *I. discreta–I. deflecta* Zone, although some of the published identifications need confirmation.

Another section of interest in a discussion of the conodont biostratigraphy across the Ordovician–Silurian boundary is on Langkawi Islands, Malaysia (Igo & Koike 1967, 1968). The latest Ordovician and earliest Silurian are represented by clastic strata ('Lower Detritus Band'), but rocks below and above this interval have produced well-preserved conodonts. Although some of their identifications need reappraisal, it appears clear that the lowest Silurian fauna recorded by Igo & Koike (1968) represents the *Pterospathodus amorphognathoides* Zone and is of late Llandovery age (Fig. 1). A modern restudy of the Langkawi succession would be of considerable biostratigraphical interest.

Changes in conodont faunas across the Ordovician–Silurian boundary

One of the most striking, if not *the* most striking, faunal turnovers during the 400 million year long history of the Phylum Conodonta occurred near the Ordovician–Silurian boundary. As recently shown (Sweet 1985: figs 7, 8), the total species diversity decreased from an estimated 75–100 species in the lower-middle Ashgill (Sweet & Bergström 1984) to about 20 species in the lower Llandovery. This diversity reduction was not a sudden catastrophic event although only a few species survived into the Silurian; rather, during the Ashgill there was a gradual disappearance involving many characteristic and long-established stocks and the new taxa that appeared were considerably fewer than those that became extinct. However, within a very limited interval, probably in the latest Ashgill, most of the remaining Ordovician taxa were replaced by forms of Silurian aspect, producing a very different appearance of the conodont faunas. From both biostratigraphical and palaeobiological points of view, it is obviously of considerable interest to establish the precise timing and detailed scenario of the conodont faunal turnover. Unfortunately, conodont data from strata reliably dated as representing the *G. persculptus* Zone, and particularly the upper part of this zone, are few and incomplete, making it currently impossible to tie the turnover closely to the graptolite zone succession. As noted below, we believe that the turnover occurred before the beginning of the Silurian (as defined by the base of the *P.*? *acuminatus* Zone), but we admit that the evidence for this conclusion is not yet conclusive. The best illustration of the faunal turnover is in the Anticosti Island succession, where there seems to be no significant stratigraphical gap in the boundary interval. As described by McCracken & Barnes (1981), the Ordovician-type conodont fauna in the Hirnantian-age Ellis Bay Formation there includes some 38 species. Immediately above a thin (0·5–2 m thick) interval having a mixed fauna, there is a Silurian-type conodont fauna of about 21 species, 16 of which are not known from older strata. Because no graptolites useful for precise zonal classification are known from the turnover interval and the immediately overlying strata, this interval cannot yet be classified in terms of standard graptolite zones and even the base of the Silurian there cannot be tied to a specific stratigraphical level.

In Fig. 4 we illustrate the known ranges of significant conodont species in the Ashgill and lower Llandovery. It should be stressed that a compilation of this type, involving data from many different sources in widely different geographical regions, will necessarily be both incomplete and probably incorrect in some respects, especially as it is based partly on arbitrary age assessments of some faunas. One interesting feature emerging from Fig. 4 is that apparently, with the possible exception of a form in the still poorly known *Ozarkodina oldhamensis* complex, not a single species with compound elements in the apparatus survived the faunal turnover. Only a few generalized species of the coniform conodont genera *Dapsilodus*, *Decoriconus*, *Panderodus*, *Pseudooneotodus*, and *Walliserodus* range into the Lower Silurian, but it should be noted that the taxonomy of some of these taxa is still not very clear.

Figure 5 summarizes the known ranges of important genera in the Ashgill and lower Llandovery. Significantly, only eight of the more than 25 Late Ordovician genera range across the turnover interval. Among these, only three (*Icriodus*, *Oulodus*, and *Ozarkodina*) have compound elements in the apparatus, whereas the five other genera have apparatuses composed of exclusively coniform elements. In our interpretation, the *Amorphognathus* lineage, which may be traced back to the Early Ordovician (Bergström 1983), became extinct in the Hirnantian. In the past, some authors have referred the early Llandovery *Pterospathodus*? *tenuis* to *Amorphognathus*, presumably on the basis of a perceived similarity in the Pa elements. However, the ramiform elements of the apparatuses of the two genera differ markedly, and we question that the Silurian species has any affinity at all with *Amorphognathus*.

The mutual relations, and possibly synonomy, of the two Ashgill genera *Birksfieldia* and *Gamachignathus* are still unclear, and it is outside the scope of the present study to discuss those matters here. However, it should be noted that it is conceivable that the ancestor of the Silurian genus *Distomodus* is to be found among this group of Late Ordovician conodonts.

The *Icriodella* lineage can be traced, with no significant interruption, from the Llandeilo to the Ashgill (Bergström 1983). We are not aware of any confirmed record of the genus in the Hirnantian but several widely distributed species have been described from the Llandovery (Aldridge 1972). The platform elements in the Silurian species are certainly similar to those in the Ordovician forms, but the non-platform elements differ in some respects, and the relations between the Ordovician and Silurian forms referred to *Icriodella* need further study; it is premature to conclude that all these forms represent the same lineage.

The Late Ordovician and Early Silurian representatives of *Oulodus* exhibit close similarity in morphology (Sweet & Schönlaub 1975) and they appear to represent the same stock. The same applies to *Ozarkodina* but this genus is not well known from the Ordovician. Its stratigraphically oldest species, *O. pseudofissilis* from the upper *A. superbus* Zone (lower Ashgill) of Britain (Lindström 1959; Orchard 1980), is isolated stratigraphically from a Hirnantian species in Scandinavia close to *O. oldhamensis*. The latter is so close morphologically to Llandovery species of *Ozarkodina* that there appears to be no doubt that they represent the same lineage.

It may be significant that the genera that survived the turnover are widely distributed in Ordovician rocks, and the species involved may have been ecologically tolerant. Sweet & Bergström (1974: 20) noted that, when known, the ancestry of most Llandovery stocks appears to be from among forms with particularly wide distribution in Midcontinent (warm-water) Province Ordovician faunas, whereas the North Atlantic (cold-water) Province stocks virtually disappeared in the Late Ordovician. A possible exception may be *Dapsilodus*, which in the Ordovician is best known from, and most common in, North Atlantic Province faunas. The severe regression reduced the space and range of environments available to the Midcontinent faunas and presumably resulted in the demise of many stocks. Many coniform taxa seem to have been less affected, particularly forms interpreted as pelagic rather than nektobenthic in habit (e.g. McCracken & Barnes 1981). The North African glaciation would have created different oceanic conditions, in terms of circulation, oxygenation and cooler temperatures. This combination of factors probably reduced the diversity of late Ashgill conodont faunas and

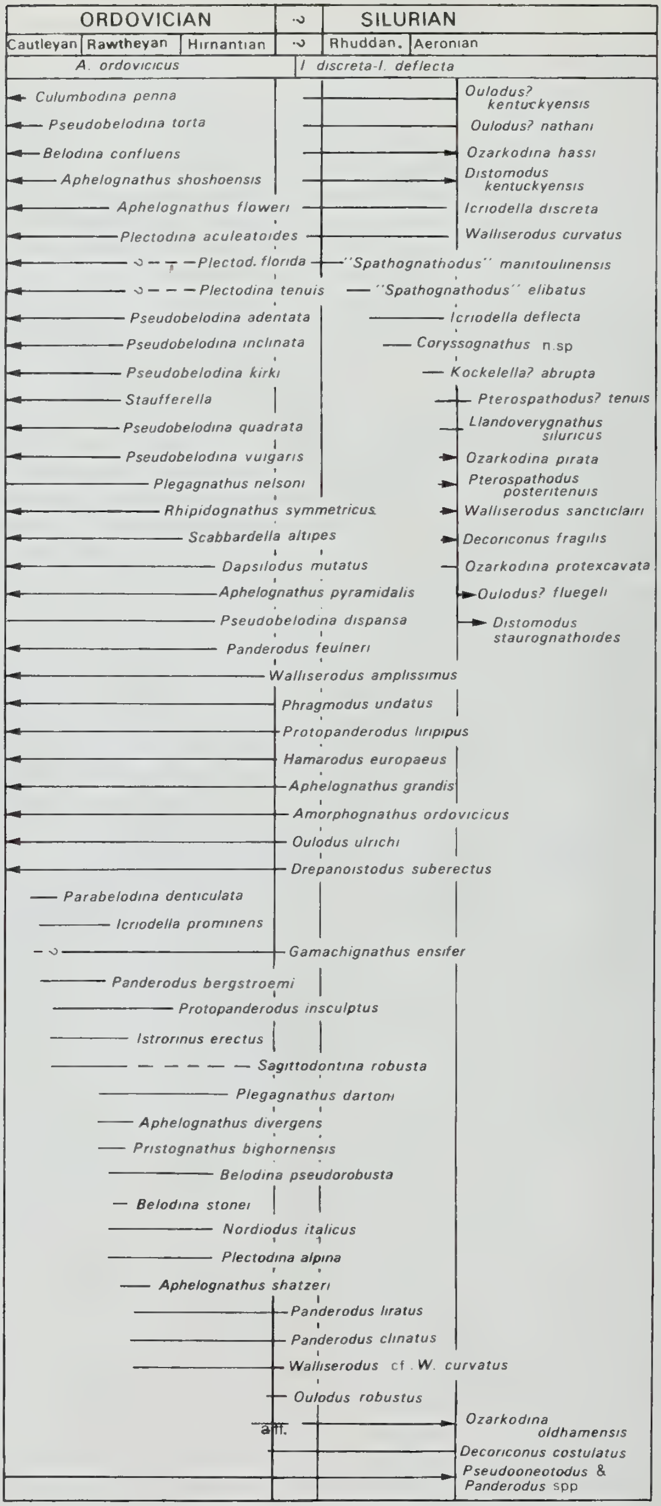

Fig. 4 Known ranges of 48 Ordovician-aspect and 22 Silurian-aspect conodont species in the uppermost Ordovician (Cautleyan to Hirnantian; *A. ordovicicus* Zone) and lowermost Silurian (Rhuddanian to lower Aeronian; *Icriodella discreta–I. deflecta* Zone). Because of uncertainties in correlations, especially in the Upper Ordovician and between continents, shown ranges are somewhat generalized. We estimate that the *A. ordovicicus* Zone contains an additional 25–50 named species that are so poorly known taxonomically and/or biostratigraphically that it is premature to include them in this diagram, and similarly a few species have been omitted from the *I. discreta–I. deflecta* Zone. Note the gradual species extinction, but also the appearance of new taxa, through the *A. ordovicicus* Zone, and the conspicuous faunal turnover in an interval near the systemic boundary.

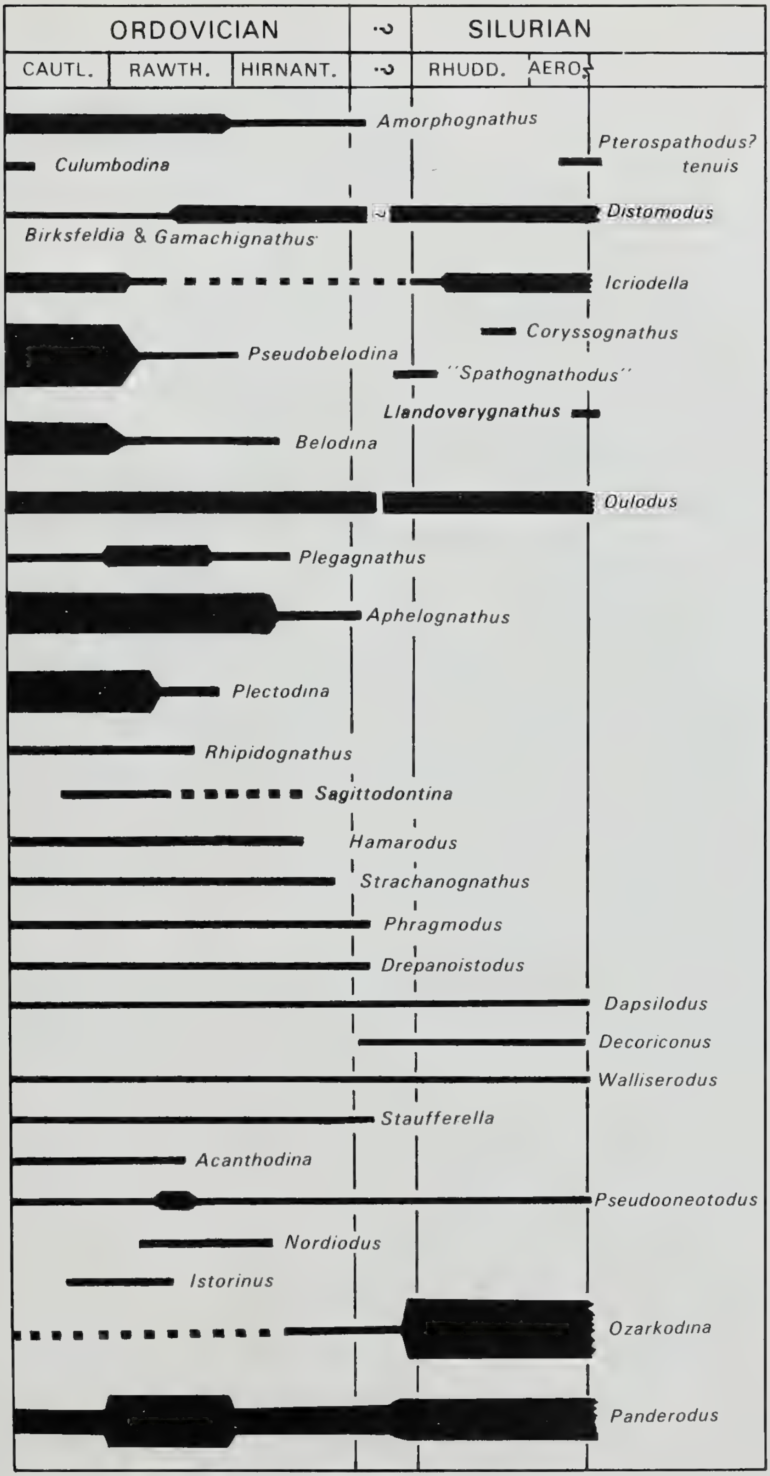

Fig. 5 Known ranges of important conodont genera in the *A. ordovicicus* and *I. discreta–I. deflecta* Zones. Relative width of vertical bars reflect number of species of a genus in a particular interval. Note that only eight of the 32 genera listed range across the systemic boundary interval, resulting in a very different generic composition of the faunas above and below.

produced the profound turnover (Barnes 1986). In addition, plate motions may have aided in these faunal changes since, by Hirnantian time, typical North Atlantic Province areas such as Baltica had moved into the tropical belt, which caused extinction of stocks long adapted to the conditions in high-latitude regions. If so, one would expect that perhaps some taxa of North Atlantic Province aspect would have survived into the Silurian in high-latitude regions, provided conditions did not become too severe. Unfortunately, high-latitude early Llandovery conodont faunas remain virtually unknown. Also, deeper-water early Silurian conodont faunas are poorly known, and it seems likely that the enigmatic origin of some of the Early Silurian platform genera may be discovered in such faunas.

Conodont Correlation of the Ordovician–Silurian boundary

As shown above, there is a profound conodont faunal change in the Ordovician–Silurian boundary interval. This faunal turnover occurs in both shallow-water cratonic and deep-water oceanic environments. The more detailed sampling and better faunal control that is feasible in carbonate platform successions is likely to provide more precise correlation within the boundary interval than can be expected in predominantly clastic deep-water oceanic deposits, which tend to contain fewer conodont-producing beds, and which are now largely preserved in structurally complex orogenic belts. Because diagnostic graptolites are largely restricted to the latter deposits, there is an obvious need to be able to recognize the systemic boundary accurately on the basis of fossils present in the cratonic successions. The geographically widespread and rapidly evolving conodonts can be expected to be helpful for precise correlations across facies boundaries also in the systemic boundary interval.

Three matters are of basic importance for the conodont correlation of the Ordovician–Silurian boundary: (1) the relation between the conodont faunal turnover and the systemic boundary in oceanic and slope sequences having zonal graptolites; (2) the relation between the conodont faunal turnover and the systemic boundary in platformal successions having key shelly fossils; and (3) if the conodont faunal turnover does not coincide with the graptolite-based systemic boundary, how do we define this boundary in terms of conodonts? All these matters involve several unsolved problems and, as will be shown below, we cannot now provide a definite answer to the last question.

In the North American platformal sequences, graptolites are rare or absent in the boundary interval. The informal units corresponding to Conodont Faunas 12 and 13, which are of Ashgill age (Fig. 1), have recently been replaced by a succession of formal conodont zones based on graphic correlation techniques (Sweet 1984: fig. 1), including the *Oulodus velicuspis*, *O. robustus*, *Aphelognathus grandis*, *A. divergens*, and *A. shatzeri* Zones. Latest Ordovician strata have a very restricted distribution on the North American craton due to a major regression associated with the Saharan glaciation(s). Further, the biostratigraphical, palaeoecological and biogeographical distributional constraints of latest Ordovician key taxa such as *Noixodontus* and *Gamachignathus* are not yet fully established. The interval of Faunas 12–13 corresponds broadly to the North Atlantic Province *Amorphognathus ordovicicus* Zone and also correlates with the lower Maysvillian to Gamachian stages, and the Cautleyan to Hirnantian stages (Fig. 1). On Anticosti Island as well as in Oklahoma–Arkansas–Missouri (Fig. 2), Gamachian cratonic faunas (Fauna 13) are associated with shelly faunas of *Hirnantia* fauna aspect. Available data show that this interval (that of Fauna 13) at least broadly correlates with the *Pacificograptus pacificus* and *Climacograptus extraordinarius* and at least part of the *Glyptograptus persculptus* Zones in the graptolite succession.

In the Yukon the first conodont faunas of Silurian aspect are found just below graptolites assigned to the *G. persculptus* Zone? (Lenz & McCracken 1982) in an interval possibly coeval with the upper part of this zone. If this interpretation is correct, it shows that the conodont faunal turnover was in latest Ordovician time and not coinciding with the systemic boundary. Lespérance (1985) has suggested that on Anticosti Island and Gaspé the level of appearance of *Acernaspis* is coeval with the base of the *P.*? *acuminatus* Zone, that is, the base of the Silurian. However, long-distance correlation of shelly fossils at the generic level is bound to be uncertain and the appearance of this trilobite in eastern Canada could obviously be younger than the base of the Silurian. On Anticosti Island, the level of the first appearance of *Acernaspis* is 30–70 m above the level of the first appearance of conodonts of Silurian aspect that mark the base of the *Oulodus*? *nathani* Zone. If the systemic boundary correlation of Lespérance (1985) is approximately correct, it is obvious that the horizon of the conodont faunal turnover is well below the base of the Silurian; it is certainly unlikely that it is higher than that stratigraphical level.

Some further data are available from other regions but unfortunately they are not decisive for establishment of the exact relationship between the conodont faunal turnover and the systemic boundary. Samples from the Carys Mills Formation of Maine and New Brunswick, possibly representing the *G. persculptus* Zone, contain faunas typical of the Silurian *Icriodella discreta–I. deflecta* Zone (Nowlan 1983; Bergström & Forbes unpublished). Conodonts of the *Amorphognathus ordovicicus* Zone are known from the *D. anceps* and *C. supernus* Zones at Dob's Linn, Scotland (Barnes & Williams, this volume) and Mirny Creek, northeast Siberia (Barnes, unpublished), respectively. Few sections are known where conodonts can be extracted from the *G. persculptus* Zone. At Mirny Creek, the basal *P.*? *acuminatus* Zone contains Silurian conodont faunas of the *I. discreta–I. deflecta* Zone, but the underlying *G. persculptus* Zone has not produced stratigraphically diagnostic conodonts (Barnes, unpublished).

We conclude that the precise correlation of the systemic boundary is uncertain in stratigraphically continuous shelly successions. Although a series of zones has been distinguished in graptolite-bearing successions, severe taxonomic problems involve several of the key species, and graptolite-based correlation into sequences with shelly fossils and conodonts is rarely possible, and conodont correlation into graptolitic facies is equally difficult. The degree of stratigraphical resolution appears greater for graptolites than for conodonts. However, Sweet's (1984) new zonal scheme for the North American Midcontinent has a resolution approaching that of the graptolite zone succession in China, and further refinements of the conodont zonal schemes are possible. If our suggestion that the conodont faunal turnover is in the upper *G. persculptus* Zone proves correct, the base of the Silurian, as now defined, will be above the interval of the most significant event in the conodont evolution of the Lower Palaeozoic. A future challenge is obviously to recover diagnostic conodonts from the *G. persculptus* Zone, and preferably from adjacent zones as well, in continuous sections, but very few sections suitable for this are known to us. In the meantime, a situation must prevail where the base of the *P.*? *acuminatus* Zone defines the base of the system in graptolitic successions, and the base of the *Oulodus*? *nathani* Zone defines a level *near* the systemic boundary in conodont sequences. Because of the prominent unconformity that is associated with the systemic boundary in most cratonic sequences, the latter level will in many, but not all, cases be the same as the systemic boundary. In stratigraphically more complete sections, it is possible that the difference between the graptolite-based boundary and the level of the conodont faunal turnover may correspond to as much as half a graptolite zone.

Conclusions

1. Although the conodont succession is known in considerable detail in both the Ashgill and the Llandovery, there are few data available from sections with rocks reliably dated by graptolites representing the upper *G. persculptus* and *P.*? *acuminatus* Zones.
2. Most boundary successions from which conodonts are known are stratigraphically incomplete or have intervals from which no diagnostic conodonts are known. The best known conodont succession across the boundary interval is on Anticosti Island, but the position of the graptolite-defined systemic boundary is uncertain there as the boundary interval lacks reliable graptolite control.
3. Ordovician and Silurian conodont faunas are strikingly different. The interval of faunal turnover is less than 2 m thick in the stratigraphically rather expanded section on Anticosti Island. The precise position, in terms of graptolite zones, of this turnover is still uncertain, but the available evidence indicates that it is likely to be in the upper part of the *G.*

persculptus Zone, below the systemic boundary. Hence it seems unlikely that the profound turnover coincides with the systemic boundary.

4. At the present time, the base of the *P.*? *acuminatus* Zone, that is the Ordovician–Silurian boundary, cannot be identified precisely on conodont evidence in sections with continuous sedimentation through the boundary interval. Further studies are needed in graptolite-controlled sections to clarify the exact relations between conodont and graptolite zones at the systemic boundary.

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Graptolite faunas at the base of the Silurian

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Synopsis

The base of the Silurian System is globally defined by the appearance of a number of species of graptoloid referable to the genera *Akidograptus* and *Parakidograptus*, as well as by a pronounced increase in species diversity from the underlying *persculptus* Zone. The nature of this diversity is given in terms of distinctive elements of the *acuminatus* Zone, in terms of its less diagnostic species, and in terms of species of local occurrence. Contrasts are made with the graptoloid faunas of the *persculptus* and *atavus* Zones.

Introduction

The ratification by the International Commission on Stratigraphy of the base of the Silurian System at the base of the *acuminatus* Zone at Dob's Linn, Scotland, greatly facilitates international correlation of the base in the graptolite facies. Even in sections with only moderately abundant or diverse graptolite associations, the *acuminatus* assemblage can usually be identified, although not always the precise lower and upper limits of the zone: the approximate correlative level is often quite clear. Furthermore, the distinctive nature of the *acuminatus* fauna makes relatively simple the present task, namely that of defining the base of the Silurian in terms of its graptolites. It should not be assumed that the base of the *acuminatus* Zone corresponds with the beginning of the post-glacial evolutionary explosion in graptoloids (Koren & Rickards 1979): that precise level is probably near the base of the *persculptus* Zone, using that term in its broadest sense. The lowest graptoloid diversity corresponds roughly with the *extraordinarius* Zone. This was followed by an increased diversity in the *persculptus* Zone, and a yet greater increase in the *acuminatus* Zone. It is the last that now identifies the base of the Silurian and which is described below.

It is helpful that the *acuminatus* Zone was originally defined at Dob's Linn (Lapworth 1878). However, he included at its base gingerbread-coloured shales which Jones & Pugh (1916) considered equivalent to Jones' (1909) *persculptus* Zone at Port Erwyd. This opinion was amply reinforced by Davies (1929), Toghill (1968) and Williams (1983), so that the original concept of the zone has been changed to mean the graptolite faunal assemblage between the *persculptus* and *atavus* Zones or their equivalents.

Graptolites immediately preceding the *acuminatus* Zone

Low diversity characterizes both the *extraordinarius* and *persculptus* Zones. There is a total absence of multistiped genera such as *Dicellograptus* and *Tangyagraptus*, and the *extraordinarius* Zone comprises only a few biserial types, including *C. extraordinarius* together with diminutive climacograptids such as *C. normalis*, *C. angustus* (=*C. miserabilis*) and *C. mirnyensis*. *C. medius* appears near the top of the zone in northeastern U.S.S.R. The *persculptus* Zone has a fauna a little more diverse than that of the *extraordinarius* Zone, but apart from rare uniserial scandent forms (*Atavograptus ceryx* and similar species) comprises biserials, including the three just listed for the *extraordinarius* Zone, but excluding *C. extraordinarius*. *Glyptograptus persculptus* itself and several closely related forms typify the *persculptus* Zone, but at least one subspecies persists into the *acuminatus* Zone. Thus the *persculptus* fauna is similar to the *extraordinarius* fauna, but differs in having the first uniserial scandent species, the very beginning of a major evolutionary explosion of these forms, and more numerous biserial species, especially glyptograptids.

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Distinctive features of the *acuminatus* Zone

The base of the zone is defined by the appearance of biserial graptolites with a characteristic drawn-out, thorn-like proximal region involving elongate sicula, elongate early thecae and a pronounced alternating arrangement of the thecal apertures. Two genera are involved: *Akidograptus* (type species *A. ascensus* Davies) with climacograptid-like thecae, and *Parakidograptus* (type species *P. acuminatus* (Nicholson)) with orthograptid-like thecae. In the lower half of the zone *A. ascensus* is usually much more abundant than *P. acuminatus*, the reverse obtaining in the upper part of the zone. However, in sections with somewhat depleted diversity the two may appear in sequence with a relatively short period of overlap. It cannot be emphasized too strongly that in richly graptolitic sections the two species seem to occur throughout, with *A. ascensus* perhaps becoming extinct a little before *P. acuminatus.*

An additional parakidograptid *P. acuminatus praematurus* was described by Davies (1929) from the lower half of the zone. Although this form has not yet been widely recorded, it has considerable potential for correlation because it is a (morphologically) earlier form than the type subspecies, having a less protracted proximal end which clearly indicates a typically more robust biserial ancestor. It is likely that *P. a. praematurus* is restricted to the lower half of the *acuminatus* Zone.

Another rare species occurring in the lower part of the *acuminatus* Zone is *Atavograptus ceryx*, although this species is more common in the *persculptus* Zone. From unpublished information and new specimens it seems likely that other, related, uniserially scandent forms will be described from this zone. Subspecies of *G. persculptus* do occur at the base of the *acuminatus* Zone, overlapping with *Akidograptus* and *Parakidograptus*, but there are also a number of other undescribed glyptograptids in both the *acuminatus* and *persculptus* Zones, often referred to as *G.* ex gr. *tamariscus*. Elucidation of these will clearly help refine correlation. *G.? avitus* extends into the lower half of the zone from the *persculptus* Zone.

C. trifilis is recorded from the middle of the *acuminatus* Zone. This tiny form has a striking three-fold spine at the base of the rhabdosome, presumably involving virgellar and antivirgellar spines. Its relationship to *C. tuberculatus* from the *persculptus* Zone is not clear; and it should be said that multispinose biserials in the Silurian are in general need of revision, as implied by Rickards & Koren (1974). *Cystograptus vesiculosus*, which lends its name to the succeeding zone in some broader zonal scenarios, occurs first of all in the upper part of the *acuminatus* Zone, as does *Climacograptus rectangularis*, a presumed derivative of the earlier *C. medius.*

Finally in this section we should mention *Orthograptus truncatus* (=*O. amplexicaulis*), *sensu lato*, which has been widely recorded from both the *persculptus* and lowest *acuminatus* Zones. The taxonomic positions of these forms are uncertain: certainly forms I recently recorded from Northern Ireland lack the proximal end spinosity of typical, earlier species, and in this sense at least are more characteristically Silurian. The same is true of Hutt's (1974) recordings of *O. t. abbreviatus.*

Less diagnostic species of the *acuminatus* Zone

The most common species in most assemblages are relatively small climacograptids which extend upwards from the Ordovician. Typical amongst these are *C. normalis* Lapworth, *C. angustus*, *C. innotatus* Nicholson and the more robust *C. medius.* In addition the diplograptids *D. modestus* and *D. diminutus* occur, the second possibly appearing in the *acuminatus* Zone, though I hesitate to claim this with the certainty the literature suggests, simply because the group is in dire need of revision. Other forms related to *C. innotatus* (sometimes referred to the genus *Paraclimacograptus*) may occur, and I have already mentioned the undescribed glyptograptids. In addition a number of sections round the world have a smaller number of forms seemingly referable to the genus *Pseudoclimacograptus* (see next section). All the forms listed in this section range upwards into the *atavus* Zone, and in some cases higher.

Species of local occurrence

In addition to the above species, modern work in several parts of the world has resulted in the recognition of what are, at present, species of relatively local occurrence. Thus *Pseudoclimacograptus* (*P.*) *orientalis* occurs in the Soviet Union, and may possibly do so in Poland (Rickards 1976: 159). In the Kolyma region Obut *et al.* (1967) record *A.* aff. *priscus* and *Orthograptus sinitzini* as well as *C. mirnyensis*. The relationship of *Orthograptus sinitzini* to *C. tuberculatus* has never been clarified and is another area worthy of further investigation, and in the recent account of the geology of northeastern U.S.S.R. (Koren *et al.* 1983) *P.* aff. *acuminatus praecedens* is recorded. Of pseudoclimacograptids Koren & Mikhailova (1983) have recorded *P. fidus* and *P. pictus*, and like forms have been found recently in the type Llandovery area (Cocks *et al.* 1984).

Waern (1948), in a careful revision of *normalis*-like climacograptids, described *C. praemedius* and *C. transgrediens*, and also recorded *C. indivisus* Davies (previously only known from the *persculptus* Zone).

The latest records from China are summarized by Mu (this volume), but it is worth noting especially that several additional records of akidograptids have been made, such as *A. xixiangensis* Yu *et al.* and *A. parallelus* Li & Jiao, as well as other biserial species as yet listed only from China. It appears correct to say that China is the only country to date with a record of the typical late Ordovician genus *Paraorthograptus* in the Silurian, i.e. in the *acuminatus* Zone. Mu (this volume) also notes the presence of several subspecies of *G. tamariscus*, but whether they are related to the later evolutionary burst of that group is not discussed.

Top of the *acuminatus* Zone

It is necessary by international agreement to define only the base of a zone. Nevertheless, it is useful here to outline what distinguishes the *acuminatus* Zone from the overlying *atavus* Zone. Basically the demise of the akidograptids is followed by increased diversification of the uniserial scandent monograptids belonging to several genera (*Atavograptus*, *Lagarograptus* and *Coronograptus*) as well as by numbers of dimorphograptids. Only in one section have akidograptids been recorded from the *vesiculosus* Zone, namely in Sardinia (Jaeger 1976). There is some overlap, naturally, but the two faunas could hardly be much more different than they are.

Finally it is clear that the *acuminatus* Zone is capable of being subdivided in useful fashion, a step already taken by Teller (1969) for example, and in effect, by Stein (1965; see also Jaeger, this volume). In most sections a lower, middle, and upper part can be identified, not only upon the occurrence of akidograptids and parakidograptids, but also on the occurrence of such species as *A. ceryx*, *C. trifilis*, *Cy. vesiculosus*, *C. rectangularis* and so on. The revision of other groups, so necessary at present, will undoubtedly increase the potential not only for international correlation at this level, but also for subdivisions of the presently defined *acuminatus* Zone.

Conclusions

The *acuminatus* fauna is not only distinctive and easily recognizable, but is widespread in the world, as the other sections in this volume make clear. The akidograptids and parakidograptids, whatever the species or subspecies, seem to be almost totally restricted to the zone. The zonal assemblage forms not only a gradual change between the *persculptus* and *atavus* Zones, but represents a distinctive stage in the evolution of Silurian graptoloids reflecting a very advanced stage of post-glacial marine transgression and the development of widespread anaerobic black shales and the re-establishment of a rich, marine, tropical plankton.

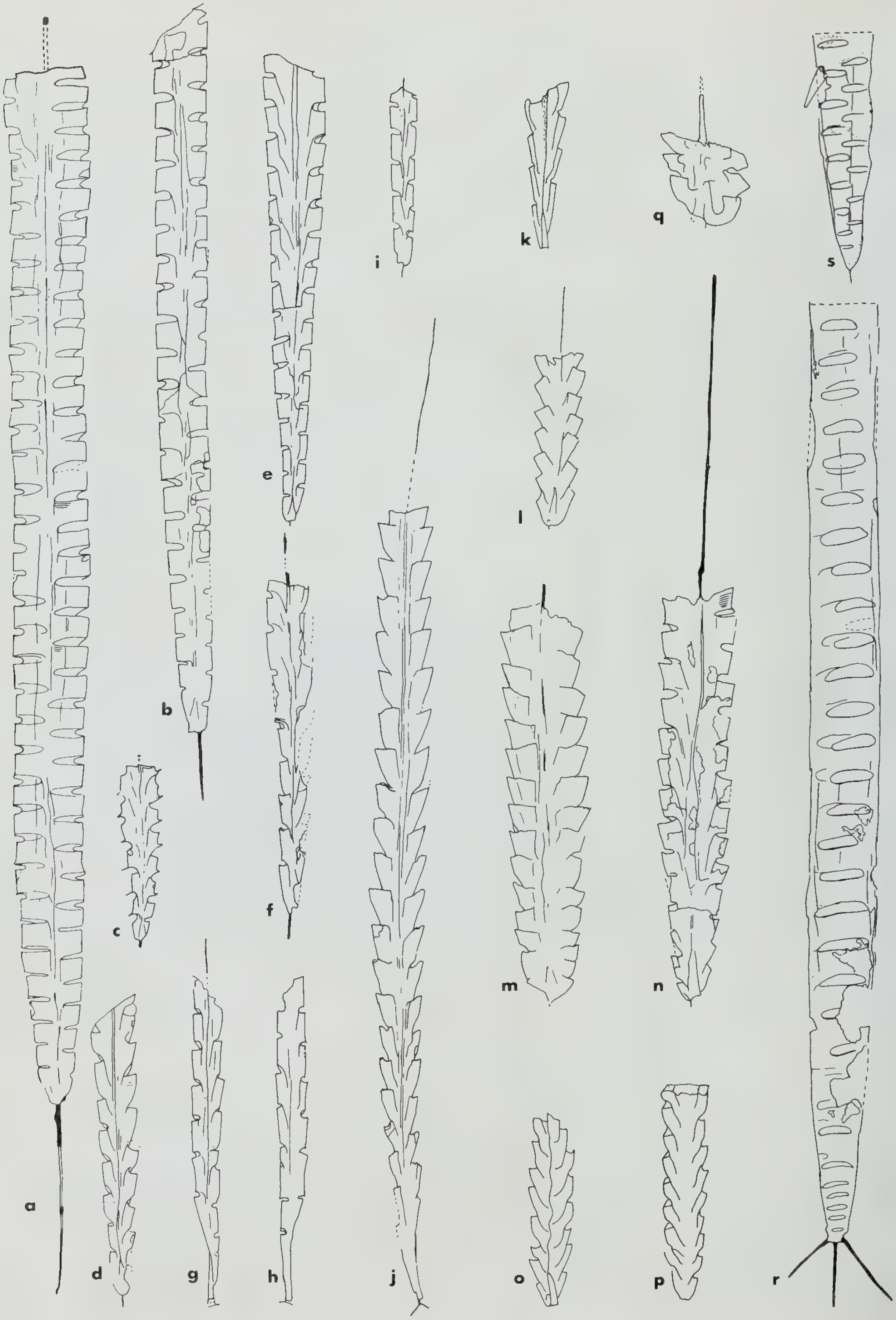
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Fig. 1 Typical *acuminatus* Zone assemblage. Specimens in Sedgwick Museum, Cambridge. a, *Climacograptus medius* Törnquist, A20150; b, *Climacograptus normalis* Lapworth, A20090; c, *Paraclimacograptus innotatus* (Nicholson), A20226; d, *Glyptograptus* sp., X.9999; e, *Climacograptus rectangularis* M'Coy, A20067; f, *Glyptograptus avitus* Davies, A10019, figd Davies, 1929: 8, fig. 21; g, h, part and counterpart of *Akidograptus ascensus* Davies, X.9996a, b; i, *Climacograptus angustus* Perner (= *C. miserabilis* Elles & Wood), X.9993; j, *Parakidograptus acuminatus* (Nicholson), A75394; k, *Parakidograptus praematurus* (Davies), A10023, figd Davies, 1929: 10, fig. 25; l, *Orthograptus* sp. (? ex gr. *amplexicaulis* Hall), X.9995; m, n, *Diplograptus modestus* Lapworth, respectively A20425 & A20428; o, p, *Glyptograptus persculptus* (Salter), *sensu lato*, figd Davies, 1929: 14, respectively figs 15 and 20 as 'mut. omega', A10013 and A10018, the latter being regarded as holotype; q, *Cystograptus vesiculosus* (Nicholson), X.9994; r, s, *Climacograptus trifilis* Manck, respectively X.9998 and biprofile view showing virgellar spine only, X.9997. All figures × 5.

Land plant spores and the Ordovician–Silurian boundary

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Synopsis

The size of early tetrad spores can be used to differentiate in a general way between late Ordovician and early Silurian rocks, although not to a fine degree of accuracy. No single trilete spores are found in Ordovician or earliest Llandovery rocks. Spores measurements are presented from the Ashgill of Bohemia, Canada and U.S.A. and the early Llandovery of U.S.A., Sweden, South Africa and Brazil.

Introduction

Early land plants can be traced through spores, having morphological analogues with spores produced by some living hepatics, back to the mid-Ordovician, about Llanvirn–Llandeilo time (Gray *et al.* 1982; Gray 1985), when recognizable remains, in terms of modern analogues, disappear. Abundant spores occur in a number of Late Ordovician (Ashgill) and in many early Silurian (Llandovery) rocks immediately above and below the Ordovician–Silurian boundary, and in some successions straddling the boundary as defined by marine invertebrates and phytoplankton. Spores occur in continental strata for this interval; they are principally abundant, doubtless related to intense weathering and often extensive metamorphism of continental rocks of this age (Gray & Boucot 1975), in shallow-water, nearshore marine rocks where other biostratigraphically useful microfossils and invertebrates are absent or inadequate for correlation. Land plant spores may ultimately prove to be the most useful fossils for helping to fix the approximate position of the Ordovician–Silurian boundary in that environment.

Gray (1985) assigned Late Ordovician–Early Silurian spores to Microfossil Assemblage Zone I. MA Zone I is a homogeneous assemblage of spores of a single morphological type: compact tetrads arranged in a tetrahedral configuration with a mean size generally less than 35 microns, and usually smooth-walled. *No* single, trilete spores are found in Ordovician or earliest Llandovery rocks, although they appear, locally, in small numbers about midway through MA Zone I. Tetrads can be assigned for the most part to *Tetrahedraletes* cf. *T. medinesis*, although this does not necessarily mean that they all represent a single taxon, since spore 'morphological species' have different taxonomic values, representing anything from families and family groups to species or subspecies. Spore tetrads are found in Late Silurian assemblages but they do not dominate in the post-Early Silurian, where they are replaced by single trilete spores, smooth-walled *and* with varied types of wall ornamentation, which find their closest morphological analogue in spores of lower vascular plants. Locally, in Ordovician–Silurian rocks from the central and southern Appalachians and the midcontinent of North America, tetrads with a reticulate surface ornamention also occur in Microfossil Assemblage Zone I, beginning in the Ashgill and continuing through the early and middle Llandovery and early part of the late Llandovery. In North America, tetrads with other ornament types appear about midway through the Llandovery (Gray *et al.* 1986: fig. 5). Tetrads with reticulate surface ornamentation have also been found in samples from Gotland, Sweden, in earliest Silurian and Ordovician–Silurian boundary rocks but have not otherwise been convincingly identified elsewhere below the Silurian, although Vavrdová (1984) claims the presence of varied ornamentations among spore tetrads from the Kosov Formation of Bohemia. I did not see these on spore tetrads extracted in my laboratory from one rock sample kindly sent to me by M. Vavrdová.

Attention has focussed on the Ordovician–Silurian boundary, and the Ashgill, a time of glaciation and widespread marine regression, as one of a small number of intervals of mass extinction among marine invertebrates and phytoplankton. Spore tetrad assemblages show no

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clearly defined changes across the Ordovician–Silurian boundary to indicate that land plants were in any way affected by the circumstances responsible for severe extinction in latest Ordovician shallow seas. There is no basic change in spore assemblages at the systemic boundary, no 'turnover' related to first or last appearances of spore types, or change in relative frequency of spore types on either side of the boundary.

The principal change that can be demonstrated for spore tetrads in Microfossil Assemblage Zone I is an increase in size from tetrads with average diameters under 30 microns in the Ordovician to tetrads with average diameters close to 50 microns near the end of Microfossil Assemblage Zone I in the mid-late Early Silurian (Gray *et al.* 1986). The consistent change in tetrad size is useful for determining the stratigraphical position within Microfossil Assemblage Zone I; change in tetrad size is less useful for discriminating the precise age of rocks to either side of the Ordovician–Silurian boundary, although tetrad size is useful for approximating the position of the boundary and for discriminating rocks close to the boundary from units of younger Llandovery age.

Spores are now known (Appendix, p. 356) from rocks deposited near the boundary from the midcontinent and Appalachians of North America; Manitoulin Island, Ontario, Canada; Brazil; Czechoslovakia; Gotland, Sweden; Libya; South Africa; and Arabia. At few of these localities is there independent information based on fossiliferous facies, shelly or graptolitic, bearing on the precise age relations of the rocks. However, marine palynomorphs (organic-walled phytoplankton: including prasinophyte phycomata and 'acritarchs') show an 'abrupt turnover' at the Ordovician–Silurian boundary related to change in phytoplankton assemblages coincident with extinction of many Ordovician species, in some southern Appalachian sections that are also spore-bearing. These have been used to position the systemic boundary in the absence of invertebrate fossils (Colbath 1983, 1985). In the absence of independent palaeontological evidence, the approximate stratigraphical position of measured spores assemblages relative to the Ordovician–Silurian boundary can be fixed, at least in North American sections, by the unconformity and lithological discontinuity at the systemic boundary itself (see Bergström & Boucot, this volume, p. 273).

Elmina Sandstone, West Africa

Spore tetrads have also been recovered from the Elmina Sandstone (lower Sekondi Series) from the vicinity of Sekondi–Takoradi, on the southwest coast of Ghana, West Africa. The Elmina was believed to straddle the Ordovician–Silurian boundary by Bär & Reigel (1980), who based their age assignment on marine phytoplankton ('acritarchs'), and in particular *Dactylofusa*, a taxon also found in strata assigned to the Itaim Formation, Maranhão (= Parnaiba) Basin, Brazil by Brito (1967: 480). Brito correlated his Palynological Zone T, from the Itaim, characterized by *Dactylofusa maranhensis*, with the Trombetas Formation of the Amazon Basin, regarded as 'probably Lower Silurian in its upper part and Upper Ordovician in its lower part' from the occurrence of *Climacograptus*, a taxon then mistakenly believed to occur only in the Lower Silurian. However, the marine, fossiliferous part of the Trombetas Formation can now be regarded as post-Lower Silurian (post-Llandovery) and probably Ludlow to possibly Gedinnian in age (Gray, unpublished spore data; P. Janvier, unpublished vertebrate data; F. Paris, unpublished chitinozoan data; L. Quadros, unpublished acritarch data 1985). Thus, Brito's assignment of Palynological Zone T from the Maranhão Basin subsurface and the coeval part of the Trombetas Formation from the Amazon Basin to the Lower Silurian–Upper Ordovician is in error. Moreover, I have recovered from the lower Trombetas, well below sections yielding marine phytoplankton, chitinozoans and vertebrates, spore tetrads of Microfossil Assemblage Zone I. Additionally Lange (1972: 38) concluded that strata from the Maranhão Basin which Brito (1967: 480) correlated with the Trombetas Formation of the Amazon Basin on the basis of shared acritarchs should be assigned to the Serra Grande Formation 'probably of Silurian age' and possibly representing lower *and* part of the middle Llandovery. Colbath (personal communication 1986) regarded the microfossil evidence provided by Bär & Reigel as inconclusive: he wrote '... they haven't illustrated any taxa which require an Ordovician age. They appear to be on safe ground in concluding that the flora is pre-Devonian, but exactly where it belongs in the Silurian is a bit tricky. The diversity of the assemblage suggests an age of approximately middle Llandovery or younger (as does the presence of *Veryhachium carminae*), but that may be an artifact of sampling... Their identification of *Dactylofusa maranhensis* appears reasonable, and does suggest correlation with the Itaim Shale in Brazil.'

Spore tetrads in the Elmina Sandstone confirm a Llandovery age assignment and indicate that the Elmina is older than Brito's Palynological Zone T in the Maranhão Basin and the marine upper Trombetas Formation in the Amazon Basin, but possibly correlative with lower Trombetas that also yields spore tetrads. The large size of the Elmina tetrads (23 (37·8) 50) based on 100 (G1473) measurements suggests mid-Llandovery rather than close to the Ordovician–Silurian systemic boundary. Finally, the sample of Elmina Sandstone collected by Bär & Reigel and later by Gray & Boucot came from a fault sliver in a badly faulted zone (all that was available). There is no assurance that this sample was near the Ordovician–Silurian boundary and there is no palaeontological evidence that requires an age near the boundary.

Manitoulin Island, Ontario, Canada

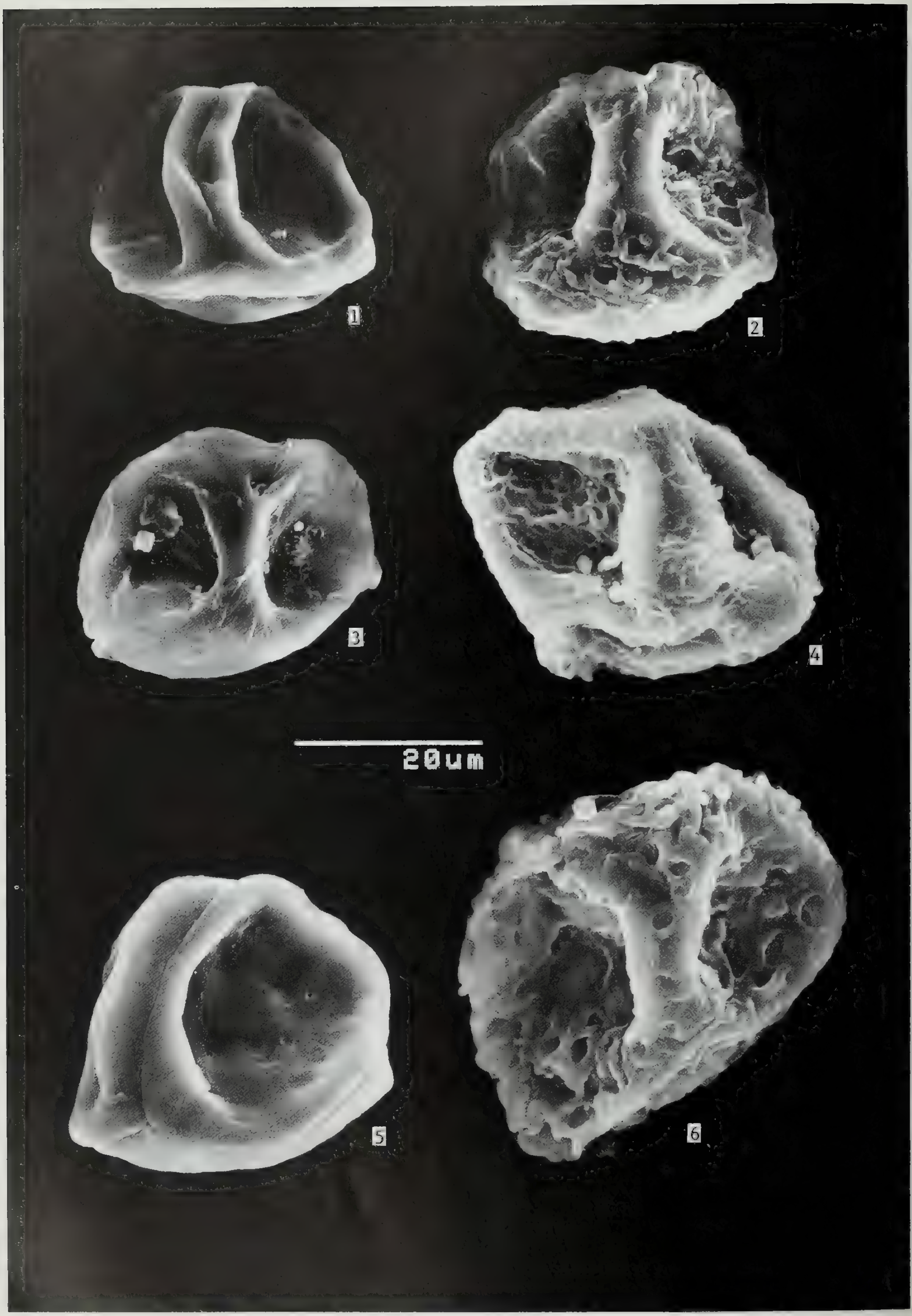
Spore tetrads come from a palaeokarst sample at, or very close to, the Ordovician–Silurian systemic boundary. The palaeokarst, represented by two surfaces, lies between the Late Ordovician (Ashgill) Kagawong beds and the basal beds of the Early Silurian (Llandovery) Manitoulin Formation on Manitoulin Island, Lake Huron, Ontario, Canada (Kobluk 1984). The boundary lies within the 0·5 m which includes the palaeokarst surfaces, but its exact position is controversial. Kobluk, who collected the samples, interprets the palaeokarsts as erosional disconformities which mark subaerially exposed surfaces that resulted from lowered sea-level at the close of the Ordovician.

Midcontinental eastern North America

Spore tetrads have been noted (Gray & Boucot 1972) in latest Ordovician–earliest Silurian beds to either side of the paraconformity that marks the boundary at Ohio Brush Creek, Ohio. Grahn & Bergström (1985: 179) have indicated, from chitinozoans, that this stratigraphical gap represents an interval from the Ashgill *Didymograptus complanatus* Zone to the early Llandovery *Climacograptus cyphus* Zone and 'hence corresponds to about four graptolite zones'—the upper Ashgill (Hirnantian or Gamachian stage) and three graptolite zones of the lowermost Llandovery. Thus the uppermost tetrad-containing Preacherville is no younger than middle Ashgill. Measured spore tetrads represent a single sample from the Preacherville Member of the Drakes Formation (called Elkhorn Formation in Gray & Boucot 1972, Gray *et al.* 1986: fig. 5) and two samples from the lowermost Silurian Belfast Member of the Brassfield Formation (G1385, G1386 from the base of the lower bed; G1384 from 10 inches above G1385 and G1386).

Eastern North America

In New York, north central Pennsylvania, southwestern Virginia, southeastern Tennessee, and northwestern Georgia various rock units to either side and encompassing the Ordovician–Silurian boundary have yielded measurable spore tetrads. These include various Llandovery formations: Whirlpool (Niagara Gorge, New York: Bolton 1957; Martini 1971; Gray & Boucot 1971), Tuscarora (Millerstown, Pennsylvania: Cotter 1982), Hagan Shale Member, Clinch (Hagen, Virginia: Miller & Fuller 1954), Red Mountain (Ringgold, Georgia: Chowns & Howard 1972), and Rockwood (Green Gap and Nickajack Dam, Tennessee: Milici & Wedow 1977). Ashgill Formations include: Red Mountain (Ringgold, Georgia), Shellmound (Nickajack Dam, Tennessee) and Sequatchie (Ringgold, Georgia; Green Gap, Tennessee). There is little independent invertebrate evidence for the age of these shallow-water, nearshore rocks to either side of the Ordovician–Silurian boundary in most of these sections and the amount of section missing at the systemic boundary may be both variable and considerable. The marked change in phytoplankton in boundary rocks reported by Colbath (1983, 1985) is the basis for positioning the boundary within a number of these stratigraphical units, including the Hagan, Nickajack Dam, Green Gap, and Ringgold Sections. Neither the Tuscarora Sandstone nor the Whirlpool (Medina Group) contains diagnostic invertebrate fossils for correlation (Berry & Boucot 1970), although field relations suggest that the lower Tuscarora, in the Millerstown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Section (Cotter 1982) and the Whirlpool, at Niagara Gorge, are early Llandovery (Gray & Boucot 1971).

Brazil

The presence of Silurian rocks in the Paraná Basin, Brazil, has long been at issue. Spore tetrads and phytoplankton (acritarchs and prasinophytes) are both consistent in suggesting a Llandovery age for the Vila Maria Formation, northeast Paraná Basin, southern Brazil, although Gray *et al.* (1985: 524) noted that the spore tetrads are similar in size to Late Ordovician *and* earliest Silurian tetrads whose average sizes are 27 to 29 microns. The Silurian age of the Vila Maria is, however, consistent with the regional geology, including the regional absence of Ordovician rocks.

Sweden

In southern Gotland, well cores at När and Grötlingbo include the entire Silurian below the Wenlock–Ludlow, based on age references provided by *Monograptus* spp., and penetrate the Ordovician–Silurian boundary; in the När core at 380·50 m (Snäll 1977). However, lowermost Silurian graptolites (*M. cometa* Zone?) are first found at 369 m (S. Laufeld, personal communication to A. Le Herisse). According to Le Herisse (personal communication) acritarch assemblages between 385·50 and 380·50 m are Late Ordovician in age, but the interval 380–372 m, characterized by red beds, is largely devoid of organic microfossils, and the 'real Silurian transgression' begins at 372 m where acritarchs and other organic microfossils are abundant. Rare spore tetrads were recovered from När samples (379, 380, 380·50, 382·50, 384 m) by A. Le Herisse, who kindly provided photographs of specimens and small splits of the cores. From three of these samples, 379 m, 380 m and 380·50 m at the Ordovician–Silurian boundary as positioned by Snäll, and 380 m, I recovered sufficient spores to measure.

Czechoslovakia

The Kosov Formation, at Hlasna Trevan near Beroun, on the Berounka River, central Bohemia, has yielded spore tetrads illustrated and described by Vavrdová (1982, 1984). The Kosov Formation corresponds to the latest Ordovician, Upper Ashgill *Glyptograptus bohemicus* Zone (Havlíček & Vaněk 1966; Havlíček & Marek 1973). Vavrdová was kind enough to provide a sample of the Kosov Formation from which abundant spore tetrads were recovered.

South Africa

Spore tetrads are known from the basal Soom Shale Member of the Cedarberg Formation, Table Mountain Group, southwestern Cape Province, South Africa. As discussed by Gray *et al.* (1986), the age of the Cedarberg Formation has been variously interpreted as latest Ordovician (Ashgill) to earliest Silurian (Llandovery) on the basis of limited invertebrate information. Cramer *et al.* (1974) bracketed the Soom Shale as latest Ordovician–earliest Silurian by chitinozoans, but favoured an Ashgill age because of brachiopod data (Cocks & Fortey 1986). Spore size is inconclusive. The measured eight samples also bracket the age of the basal Soom Shale as latest Ordovician–earliest Silurian. However, J. N. Theron's recent discovery of conodont assemblages there, considered to be of late Ordovician age by a number of specialists, confirms an Ashgill age for the unit.

Conclusions

These preliminary results, with size frequency measurements, show that the Ordovician–Silurian boundary is bracketed by spore assemblages with spore tetrads having average sizes

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Figs 1–6 Scanning electron micrographs of obligate tetrahedral tetrads of spores typical of Microfossil Assemblage Zone I (Gray 1985). Magnification × 1500. All from the Ashgill Preacherville Member, Drakes Formation, Ohio Brush Creek Section, Kentucky, U.S.A. (G1285). Most spore tetrads from Microfossil Assemblage Zone I are smooth-walled (Figs 1, 5), and some have an outer envelope that may be shed. The outer envelope is most commonly reticulate (Figs 2, 4, 6). Fig. 3 shows a spore tetrad with a smooth-walled envelope or possibly a degraded reticulate envelope.

less than 30 microns. The average size of spore tetrads to either side of the boundary, as positioned by palaeontological or micropalaeontological data, or by a stratigraphical gap and change in lithology, is about 26–29 microns, although there are both smaller (Sequatchie Formation) and larger spore tetrads (Manitoulin palaeokarst) known from rock units close to, or at, the systemic boundary. Slight differences in spore tetrad size on opposite sides of the boundary are inadequate, without other evidence, to distinguish latest Ordovician from earliest Silurian age rocks, although the Ordovician–Silurian boundary is easily bracketed by spore assemblage measurements.

I have no explanation for the relatively small size of the spore tetrads from the Sequatchie Formation. The measured samples may be lower in the Sequatchie, i.e. older, than now recognized in terms of their stratigraphical position relative to the Ordovician–Silurian boundary, possibly related to the presence of a significant disconformity. I have no independently dated assemblages from within the Ashgill for comparative purposes. With small microfossils, there is always the possibility of independent size-sorting, since these fossils behave as clastic sedimentary particles with hydraulic equivalents in the fine or very-fine silt size fraction (Stanley 1969; Muller 1959; Brush & Brush 1972). Water turbulence can keep large quantities of pollen or spores in suspension for extended periods, and it may be that the smaller spore tetrads of the Sequatchie were winnowed from the spore assemblage through progressive sorting and deposited with finer mineral particles, possibly in a more off-shore environment than represented by the depositional environments of many of the other units, or in a pattern related to marine currents or some other hydrodynamic factors. This phenomenon may also account for some of the inconsistencies found in a few of the other measurements. The large size of the Manitoulin tetrads is not consistent with the other results and a more serious threat to the utility of spore-size measurements for discriminating the Ordovician–Silurian boundary, since the stratigraphical position of the sample seems well fixed. The comparatively large size of these spores, for which only relatively few measurements were available, may reflect the fact that this sample was not originally extracted for spores, but for arthropod cuticle remains, so that smaller tetrads may have been lost in the sieving process. This material is being re-extracted specifically to recover spores and measurements repeated on a larger number of spore tetrads.

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Appendix

Size measurements of Ashgill and Early Llandovery spore tetrads

Lower Llandovery	N	Min.	Aver.	Max.
Robert Moses Power Plant Section, Niagara Falls, New York				
Whirlpool Sandstone (G1189)	250	13	26·5	44
Millerstown Section, Pennsylvania				
Tuscarora Formation (G1408)	100	18	27·0	41
Tuscarora Formation (G1407)	100	17	27·3	47
Tuscarora Formation (G1406)	100	17	28·0	49
Tuscarora Formation (G1374)	150	16	27·5	45
Nickajack Dam Section, Tennessee				
Rockwood Formation (ND70)	41	22	33·6	51
Rockwood Formation (ND54)	107	15	29·9	53
Ringgold Section, Georgia				
Red Mountain Formation (RN570)	200	15	26·6	39
Red Mountain Formation (RN470)	86	17	29·4	48
Red Mountain Formation (RN420)	45	17	27·8	39
Red Mountain Formation (RN370)	148	13	25·6	38
Red Mountain Formation (RN320)	98	13	26·9	45

Hagan Section, Virginia	N	Min.	Aver.	Max.
Hagan Shale Member, Clinch Formation (HGII70)	200	13	29·2	49
Hagan Shale Member (HGII50)	135	19	32·2	54
Hagan Shale Member (HGII30)	87	18	29·7	49
Hagan Shale Member (HGII10)	215	13	27·5	47
Ohio Brush Creek Selection, Ohio				
Belfast Member, Brassfield Formation (G1384)	100	17	26·9	39
Belfast Formation (G1385)	100	18	27·3	45
Belfast Formation (G1386)	150	17	27·0	40
Narborrningen 1, southern Gotland, Sweden				
Unnamed formation, 379·00 m (G1553)	25	19	28·8	41
Unnamed formation, 380·00 m (G1549)	69	20	28·6	52
Unnamed formation, 380·50 m (G1548)	34	19	29·8	40
Fazenda Tres Barras Section, Brazil				
Vila Maria Formation (G1391)	150	18	29·1	42
Ashgill				
Swartleikloff Section, South Africa				
Soom Shale Member, Cedarberg Formation (G1363)	100	15	27·5	40
Soom Shale Member (G1364)	100	17	28·5	37
Soom Shale Member (G1365)	100	20	28·4	41
Soom Shale Member (G1366)	100	17	27·5	39
Soom Shale Member (G1367)	100	22	30·5	40
Soom Shale Member (G1368)	100	17	28·7	40
Soom Shale Member (G1369)	100	20	29·6	43
Soom Shale Member (G1370)	100	17	29·2	45
Combined average	800	15	28·8	45
Hlasna Treban Section, Bohemia	108	16	28·0	47
Kosov Formation (G1430)				
Paleokarst at Ordovician–Silurian systemic boundary, Manitoulin Island, Ontario (G1272)	45	22	33·0	46
Ohio Brush Creek Section, Kentucky				
Preacherville Member, Drakes Formation (G1285)	252	17	27·5	53
Green Gap Section, Tennessee				
Sequatchie Formation (GG19)	58	12	22·0	32
Nickajack Dam Section, Tennessee				
Shellmound Formation (ND33)	150	18	26·5	50
Shellmound Formation (ND20·5)	141	16	27·7	43
Ringgold Section, Georgia				
Red Mountain Formation (RN210)	59	16	25·6	41
Red Mountain Formation (RN201)	89	14	23·9	46
Sequatchie Formation (RN195 = G1245)	200	11	23·2	40
Sequatchie Formation (G1245)	100	11	24·0	46
Sequatchie Formation (G1246)	100	14	23·7	50
Sequatchie Formation (RN139)	66	14	24·4	35

Notes: The samples are in stratigraphical order within each section, with the youngest at the top. G numbers are Gray extractions; others are Colbath extractions. G1385, G1386 were measured from samples collected along the strike.

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Trilobites

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Synopsis

Hirnantian (latest Ordovician) trilobite faunas are surveyed. Some are of restricted diversity, but others are highly diverse. A coeval trilobite fauna from the Gamachian Stage of Anticosti Island is highly diverse, but of different zoogeographical affinity. *Dalmanitina–Mucronaspis* occurrences, of putative Silurian age, and usually with other shelly fossils, are discounted. The base of the *acuminatus* Zone may correlate with the first appearance of the trilobite *Acernaspis* in shelly sequences but this awaits confirmation. The systematics of spinose hypostomata within the Dalmanitidae are critically examined, and it is concluded that the concept of *Mucronaspis* requires that spinose hypostomata be present before the generic name is applied. A lectotype of *Mucronaspis danai* (Meek & Worthen 1866) is selected. *Dalmanitina* (*Songxites*) sp. (nov.) from Dob's Linn and *Mucronaspis danai* from Illinois and Missouri are illustrated and described. *Mucronaspis mucronata* and *Acernaspis norvegiensis* are also illustrated, but only briefly discussed. *Acernaspis* (*Acernaspis*) *salmoensis* sp. nov. from Anticosti Island and *Cryptolithus portageensis* sp. nov. from Percé are erected.

Introduction

The correlation of the base of Silurian, as defined at Dob's Linn, Scotland, using trilobites is difficult as major changes in trilobite faunas occur near, or at, this boundary. In the following, the term Hirnantian (stage) will be used for the strata immediately underlying the *acuminatus* Zone, while the Rhuddanian is the oldest Silurian stage; the Hirnantian, however, has not been approved by the International Union of Geological Sciences.

The disappearance of many trilobite genera and families in the latest Ordovician is well known. Thomas *et al.* (1984: 39) noted that the change from the Ashgill stages Rawtheyan to Hirnantian, in England and Wales, entailed the disappearance of many genera and important Ordovician families such as the agnostids, Trinucleidae, Remopleurididae, Telephinidae, Cyclopygidae, Asaphidae, Dionididae and Phillipsinellidae. Trinucleidae are now known to extend into the Hirnantian (see below). Asaphidae (from Scotland) and Cyclopygidae (from Ireland) were, however, reported from the Hirnantian by Thomas *et al.* (1984: 41, 44). The Hirnantian Stage is reputed for its distinctive impoverished trilobite faunas (see also Lespérance 1974), although the degree and nature of impoverishment is variable from region to region. It would thus appear from these and other data that the major trilobite extinction was near the Rawtheyan–Hirnantian boundary, and not at the base of the Silurian.

Lespérance (1985) attempted to correlate the base of the *acuminatus* Zone with shelly sequences. He noted an ordered succession of appearances of faunas and taxa on Anticosti Island and elsewhere: the *Oulodus*? *nathani* (conodont) Zone, followed upward by the brachiopods *Zygospiraella*, succeeded by *Stricklandia*, then the trilobite *Acernaspis*, and finally the brachiopod *Virgiana*. Only the appearance of *Acernaspis* seemed to coincide with the base of the *acuminatus* Zone, when compared with the Oslo region (Norway) and the USSR (Kazakhstan and northeast USSR). This *acuminatus–Acernaspis* correlation has still to be further tested and confirmed, but no additional data have since come to light to contradict or reaffirm it; it is therefore accepted and used herein.

The recognition of trilobite faunas immediately younger than the base of the *acuminatus* Zone is exceedingly difficult if one excludes *Acernaspis*. Trilobite genera recorded from lowermost Silurian (Rhuddanian) strata consist of holdovers from the Ordovician, and show little change from their ancestors. This apparent lack of change may, however, be due more to the scarcity of monographic treatment, poor preservation and/or, more probably, to infrequent

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preservation, than to lack of evolution. *Stenopareia*, aulacopleurids, proetids and calymenids, although apparently common, seem to show little change, or, at least, stratigraphically useful species have not been recognized. Lichids and odontopleurids are scarcer, but still widespread; again stratigraphically useful species are not evident. Homalonotids are even scarcer. All these lowermost Silurian taxa should be reviewed in the light of new material.

Rhuddanian trilobite faunas are perhaps notable by the presence of a limited number of Ordovician holdovers. Examples are *Cyphoniscus* cf. *socialis* (Salter 1853) associated with *Acernaspis* (*A.*) *primaeva* (Clarke 1908) and other trilobites in the Matapédia Group north of Percé (Lespérance *in* Ayrton *et al.* 1969: 476; Dean 1972), and *Hadromeros* which has been widely reported lately in Rhuddanian strata. Lane (*in* Thomas *et al.* 1984: 53) reports the presence of *Panarchaeogonus* and *Ceraurinella* in the later Llandovery, so that these otherwise typical Ordovician genera must also have been present in the Rhuddanian. *Sphaerocoryphe* has also been reported from an unspecified level in the Silurian (Thomas & Lane 1984: 62). Thomas *et al.* (1984: 52) state that the following genera are unknown from the Ordovician: *Warburgella* (*Warburgella*), *Harpidella* (*Harpidella*), *Dalmanites*, *Anacaenaspis*, *Podowrinella*, *Calymene* s.s. and *Acanthopyge* (but see below). All in all, early Silurian trilobite faunas appear to be characterized by the absence of specialized Ordovician families and genera, and by the presence of 'generalized' forms, rare new ones (notably *Acernaspis*), and some holdovers from the Ordovician. The 'generalized' trilobites yield in many instances (in the later Silurian and Early Devonian) specialized and distinctive descendants. The early Silurian trilobite faunas thus stand between distinctive and specialized faunas, both older and younger.

This contribution will consequently focus on a certain number of biostratigraphically useful taxa which were abundant, or at least well known, in the latest Ordovician or earliest Silurian.

Hirnantian and Gamachian trilobite faunas

Lespérance (1974) surveyed Hirnantian brachiopod and trilobite faunas. Some of this is still pertinent, but must be viewed in the light of the recently promulgated *acuminatus* boundary. Subsequent data from the midcontinent of the USA (Amsden 1974), China (Nanjing Institute 1984), Wales (Cocks & Price 1975; Cocks *et al.* 1984), Norway (Brenchley & Cocks 1982), and the USSR (Apollonov *et al.* 1980; Koren *et al.* 1983) have since been added.

Precise correlations of shelly faunas near and at the *acuminatus* boundary are hampered by the lack of continuous thoroughly investigated sections possessing enough elements in common to correlate. The Anticosti (and, accessorily, Percé) and Oslo region sections are at present those that are easiest to correlate, and they permit, in turn, additional correlations with other sections. The basal *Oulodus*? *nathani* Zone occurs in the lower part of member 7 of the Ellis Bay Formation on Anticosti, and this zone also occurs very low in the Solvik and Sælabonn Formations of the Oslo region (Worsley 1982; Lespérance 1985). It is inescapable that the ecologically complex and diverse faunas of the latest Ordovician '5a' and '5b' of the Oslo region (Brenchley & Cocks 1982) must correlate with strata below the lower part of member 7 on Anticosti. Only '5b' (Langøyene and Langåra Formations) is Hirnantian, whereas the lower boundary of the Gamachian (at the base of the Ellis Bay Formation) is older than the base of the Hirnantian (it occurs 34 m above the base of the 130 m thick Burmingham Member in the Percé area, Lespérance, this volume). To compare Hirnantian faunas on Anticosti and the Oslo region, it is necessary to draw the base of the Hirnantian within the Ellis Bay. As no drastic drop in diversity is apparent in the Ellis Bay (as present in the type Rawtheyan–Hirnantian), quite to the contrary, members 1 and 2 are arbitrarily excluded from the following discussion (representing a thickness comparable in proportion to the Percé strata). Pre-*Oulodus*? *nathani* Zone trilobites common to Anticosti and the Oslo region are *Platycoryphe* and *Toxochasmops*. *Calyptaulax*, *Decoroproetus*, *Dicranopeltis*, *Harpidella*, *Illaenus*, *Mucronaspis*, *Panderia* and *Stenopareia* are only known from '5b', whereas *Amphilichas* (two species), *Cyphoproetus*, *Erratencrinurus* (*Celtencrinurus*), *Failleana*, *Hemiarges*, *Isotelus*, *Lichas*, *Nahannia*, *Otarion*, *Paraharpes* and *Sphaerocoryphe* are only known from Anticosti (Bolton 1981; Brenchley &

Cocks 1982; Chatterton *et al.* 1983; and the writer's unpublished data). To compare post-basal *Oulodus*? *nathani* and pre-*acuminatus* trilobite faunas from the same areas, all '6a' and '6b' occurrences are presumed to predate the first occurrence of *Acernaspis* (this is probably too generous, as it first occurs in the upper half of '6bα') (data are from Chatterton *et al.* 1983; Helbert *et al.* 1982; and the writer's unpublished data). *Cyphoproetus, Diacalymene, Harpidella* and *Stenopareia* occur in both areas, but *Amphilichas, Astroproetus, Failleana, Illaenoides, Leonaspis* and *Primaspis* occur only on Anticosti, while *Arctinurus, Calymene, Dicranopeltis* and *Hadromeros* only in the Oslo region.

From the above survey, it is clear that there are few Hirnantian genera in common between Anticosti and the Oslo region, which suggests significant zoogeographical differences. If one tabulates the genera restricted to either region, throughout the whole Hirnantian, 23 are counted. Of these, 9 can be considered long-ranging, and 11 seem to be typical Ordovician genera at the end of their biozones (*Amphilichas, Calyptaulax, Erratencrinurus* (*Celtencrinurus*), *Failleana, Illaenus, Isotelus, Mucronaspis, Nahannia, Panderia, Paraharpes, Primaspis*). The remaining three (*Arctinurus, Calymene* and *Illaenoides*) are more typical of the Silurian, and their biozones should consequently be extended downwards. The genera common to both in pre-*Oulodus*? *nathani* strata are typical Ordovician ones, while those common to both in post-*Oulodus*? *nathani* strata are long-ranging.

As the nearby Percé area was assuredly on the same platform as Anticosti and it has very little in common with Anticosti (or the Oslo region), one must seek an explanation. The most obvious reason for these differences is ecological control on these faunas, and, particularly, depth of water and temperature. Depth, *per se*, appears insufficient to explain these differences. Water temperature, particularly considered with an upward-moving thermocline (and glaciations?), appears far more plausible an explanation for these zoogeographical differences.

Finally, what does a typical Hirnantian trilobite fauna contain? Benthic Assemblage 6 faunas consist wholly or predominantly of trilobites, and can be composed of few or many taxa, but shallower communities have far fewer trilobites, commonly with abundant brachiopods. Excluding for the purpose of this discussion groups other than trilobites, two distinct trilobite faunas apparently coexisted. A North American type appears evident (Anticosti Island, Ellis Bay Formation; other faunas such as the Mackenzie faunas reported by Chatterton & Ludvigsen 1983, but sparingly developed in view of the profound disconformity between the Ordovician and the Silurian in most places in North America). The typical 'Old World' Hirnantian trilobite fauna can be monospecific to highly diverse, but usually includes *Dalmanitina* or *Mucronaspis* and a homalonotid (*Brongniartella* or *Platycoryphe*). The Oslo region faunas appear to be intermediate between the two. On the other hand, this variation in diversity has also been ascribed to nearness to the center of glaciation (Cocks & Fortey 1986), but the problem appears more complex than that explanation suggests.

Dalmanitina–Mucronaspis taxa near the Ordovician–Silurian boundary

A bewildering number of species, particularly from China, and variously referred to *Dalmanitina* or *Mucronaspis*, have been reported from strata immediately above or below the previously accepted or assumed Ordovician–Silurian boundary. Apart from the difficult systematics associated with the generic assignment of the various species (a few are discussed at some length below), some of them have been taken as indicative of a Silurian age. These putative Silurian species are: *Mucronaspis danai* (Meek & Worthen 1866), *Dalmanitina hastingsi* (Reed 1915), *D. kosyndensis* Balashova 1966*, *D. malayensis* Kobayashi & Hamada 1964, *D. brevispina* Temple 1952, *D. nanchengensis* Lu 1957, *D. pamirica* Balashova 1966* and *D. subduplicata zorbata* Balashova 1966* [*: as cited by Kobayashi & Hamada 1971, but primary source unverified by the present author].

It will be shown below that these occurrences are logically assigned to the Ordovician, if one accepts the base of the Silurian as *at the first appearance* of the *acuminatus* Zone. This principle of correlation by first appearances is at the heart of recent stratigraphical practice, and underlies the choice of 'golden spikes', as exemplified by the choice of the Silurian–Devonian boundary. If this is followed, strata underlying the *acuminatus* boundary *must* be assigned to the Ordovician, whatever the sedimentological and/or faunal succession may suggest.

The primary types of *Mucronaspis danai* occur in an erosional channel, assigned to the Leemon Formation, within the Girardeau Limestone of southern Illinois. Conodonts within the same beds as the trilobite are of the *Amorphognathus ordovicicus* fauna (Thompson & Satterfield 1975), of undoubted Ordovician age. Whether this occurrence is of Richmondian or Gamachian age is unknown. The species also occurs in the Edgewood Group of northeastern Missouri (see below).

Dalmanitina hastingsi occurs in the lower, or trilobite, unit overlain by the upper or graptolite unit, of the Panghsa-pye Formation (Bender 1983: 63) in Burma. This lower unit is only known from the Panghsa-pye region itself, where it is underlain by the Nyaungbaw Limestone, which is Late Ordovician on the basis of conodonts (Wolfart *et al.* 1984: 41). The graptolites from the upper Panghsa-pye have been assigned to the Rhuddanian (but not as old as the *acuminatus* Zone). The brachiopods from the lower trilobite unit are closely related, if not identical in many cases, to Hirnantian forms (Temple 1965). There is thus no compelling evidence to consider *D. hastingsi* Silurian, and it is here assigned to the Ordovician.

Dalmanitina malayensis occurs 1·4 to 1·8 m above the base of the Detrital Band in the Langwaki Islands, above graptolites (Kobayashi & Hamada 1971, 1974) of the *persculptus* Zone. The topmost 4·7 m of the 25 m thick Detrital Band yields graptolites of the upper Rhuddanian–Aeronian. There is consequently no reason to consider *D. malayensis* Silurian.

The primary types of *Dalmanitina brevispina* originate from Watley Gill (Lake District of northern England), from a limestone of the 'Silurian Basal Beds'. Graptolites of the *acuminatus* Zone are welded (*sic*) on top of the 'Basal Beds' (Rickards 1970: 7). There is no evidence for such a zonal assignment for the 'Basal Beds', or strata below them. The same species occurs at Keisley, where it is known from strata below the *persculptus* and *acuminatus* Zones (Wright 1985). Thus both the Keisley and Howgill Fells occurrences of *D. brevispina* are probably Ordovician.

The type material of *Dalmanitina nanchengensis* comes from southern Shaanxi, and it occurs above beds yielding the graptolites *Climacograptus angustus* (Perner, 1895) and *C. mirnyensis* (Obut & Sobolevskaya, 1967) (Lu & Wu 1983). Although *D. nanchengensis* is also known from Sichuan–Guizhou (Szechuan–Kweichow), it is the Shaanxi occurrence that is considered Silurian, on the basis of *C. mirnyensis* which apparently occurs only in the *acuminatus* Zone. Koren *et al.* (1983), however, report that *C. mirnyensis* occurs in the *extraordinarius*, *persculptus* and *acuminatus* Zones, so that *D. nanchengensis* is herein assigned to the Ordovician, because of the lack of diagnostic Silurian elements below it.

Apart from these species, *Mucronaspis mucronata* (Brongniart 1822) has also been claimed to occur in Silurian strata. Disregarding the Scandinavian claims to this age, which are now abandoned in Scandinavia itself, *M. mucronata* has been so cited in the Percé area and in Kazakhstan. Lespérance (this volume) assigns the Percé occurrences to the Hirnantian, while the Kazakhstan occurrences (which cannot be proven to belong to *Mucronaspis*), with other shelly faunas, are in the *persculptus* Zone (Apollonov *et al.* 1980) and so they are pre-Silurian.

Dalmanitina sp. occurs in the '*Protatrypa*' assemblage, which may reach a level as high as the *Coronograptus cyphus* Zone (Mu 1983: 116–7) in China. In accord with Williams (1983: 611), the base of the *acuminatus* Zone in China is higher than elsewhere, and hence the *Dalmanitina* sp. is perhaps largely pre-Silurian in age; stratigraphical details are not sufficient for a more extended discussion.

The Haverford Mudstone Formation of Wales has yielded in its lower 235 m '*Mucronaspis mucronata*' (quotes are this writer's) and other fossils (Cocks & Price 1975), assigned to the Hirnantian, while the uppermost 140 m yields a rich Rhuddanian fauna, containing, *i.a.*, *Acernaspis* sp. *Brongniartella* sp., *Hadromeros elongatus* (Reed 1931) and *Dalmanites* sp. (Temple 1975); the generic assignment of the dalmanitacean is noteworthy, as are its associated trilobites.

Systematic Palaeontology

Family **DALMANITIDAE** Vogdes, 1890

The distinction between the genera *Dalmanitina* and *Mucronaspis*, as well as the proper assignment and distinctive characters of the numerous species referred to these genera, is difficult. The most recent treatments are by Ingham (1977); Lespérance & Sheehan (1981); Owen (1982); Lu & Wu (1983); Zhu & Wu (1984); Wu (1984); and Cocks & Fortey (1986). Zhu & Wu (1984: 89) were uncertain whether a denticulate posterior hypostomal margin was diagnostic of *Mucronaspis* and, if so, no genuine *Mucronaspis* would be present in China. Hypostomata are conservative evolutionary features and, potentially, powerful phyletic tools, which is a truism in trilobite systematics. As both Destombes (1972) and Ingham (1977) stressed the presence of a denticulate (spinose) hypostoma in *Mucronaspis*, a survey of Ordovician dalmanitacean hypostomata is instructive.

Llanvirn spinose hypostomata are unknown. Three are known from the Llandeilo: *Eodalmanitina macrophtalma* (Brongniart, 1822) (the type species of the genus, Henry 1965: pl. 6, fig. 2), *Crozonaspis struvei* Henry, 1968 (Henry 1980: 149) (but *Crozonaspis morenensis morenensis* Hammann, 1972 (Hammann 1974) is not spinose), and *Phacopidina micheli micheli* (Tromelin, 1877) (Henry 1980: 128). These hypostomata have two small spines (or 'denticles') on their posterolateral border. Caradoc spinose hypostomata also have two spines or denticles: *Kloucekia* (*Phacopidina*) aff. *solitaria* (Barrande, 1846) (of Destombes 1972), *Mucronaspis zagoraensis* Destombes, 1972 (but hypostoma not illustrated), *Dalmanitina* (*Dalmanitina*) *socialis* (Barrande, 1846) (of Struve 1958: pl. 2, fig. 14), the one questionably referred to *Eudolatites* cf. *angelini* (Barrande, 1852) by Struve (1958: 208; pl. 2, fig. 11), as well as the upper Caradoc and Ashgill *Baniaspis globosa* Destombes, 1972. The following Ashgill spinose hypostomata have six spines: *Mucronaspis danai*, *Dalmanitina* (*Mucronaspis*) *termieri* Destombes, 1963 (the type species of the subgenus), and *Mucronaspis mucronata* (Brongniart, 1822). Except for *Crozonaspis*, and the aff. *solitaria* of Destombes (1972; see below), the genera appear to be characterized by these spinose hypostomata, but the hypostomata of most named species are unknown.

The hypostoma of *Dalmanitina mucronata* illustrated by Kielan (1960: pl. 20, fig. 6) is spinose, but it is uncertain if two or six spines are present. Ingham (1977: 113; pl. 25, figs 3–4) described a small holaspis of *Mucronaspis mucronata* which has marginal denticles; he compared this specimen with Kielan's (1960) illustration. Here again, it is not clear how many spines are present; additional data are needed on these unique (?) Polish and northern English occurrences. *Eudolatites* (*Deloites*) *maiderensis* Destombes, 1972 (the type species of the subgenus) is said to have the beginnings of three small 'denticles', from a worn posterior border of the hypostoma; again more data are needed to confirm this unique type of spinosity. These three occurrences are apparently all Hirnantian.

From the spinose hypostomata previously enumerated, five appear to share common traits: significantly greater width than length (ratio as 4:3), essentially identical shapes (strongly curved posteriorly, lateral margins subparallel), a distinct lateral and posterior border, with two or six denticles or spines. These five are: *Crozonaspis struvei*, *Eodalmanitina macrophtalma*, *Kloucekia* (*Phacopidina*) aff. *solitaria* of Destombes 1972, *Dalmanitina* (*D.*) *socialis* of Struve 1958, and *Mucronaspis termieri*. However, significant nomenclatorial problems exist with two of the above taxa. The lectotype of *Sokhretia solitaria* (Barrande, 1846) (the type of the genus) has been illustrated (Šnajdr 1982), and it is obvious that it is not conspecific with the Moroccan species. This Moroccan aff. *solitaria* falls within the concept of the genus *Phacopidina* of Henry 1980, and is consequently better referred as *Phacopidina* n. sp. The second nomenclatorial problem is, however, far more serious. Barrande's (1852: pl. 26, fig. 21) illustration of the hypostoma of *Dalmanitina socialis* (the type of the genus) shows no denticles, and Struve's (1958) illustration of the species appears to differ only in the presence of these hypostomal denticles. Either hypostomata are sexually dimorphic, they are phenotypically variable, or significant parallel evolution exists within the Dalmanitidae, with consequent polyphyly. Parallel evolution appears much more plausible to this writer, if only to explain the notoriously difficult systematics associated with some dalmanitaceans. If this explanation is correct, it also necessitates a revision of many previously held taxonomic concepts. Be that as it may, Struve's (1958) *socialis* is better called *Mucronaspis* sp. (nov).

Denticles on hypostomata apparently appeared in the Llandeilo; originally two in number, Ashgill representatives acquired six. Some denticulate hypostomata do not fit into the five taxa quoted above, and one is led to conclude that a separate branch diverged in the Caradoc. These considerations indicate that denticles, or spines, are diagnostic of the hypostomata of *Mucronaspis*, if only because a possible evolutionary path leads to it. If this is the case, the numerous Hirnantian species which are problematically assigned to *Dalmanitina* or *Mucronaspis* should accord with what the type species of the two genera in question possess: non-denticulate in *Dalmanitina*, and denticulate (or spinose) in *Mucronaspis*. Other generic characters of *Mucronaspis* (as opposed to *Dalmanitina*) have been given by Ingham (1977) and Owen (1982). *Mucronaspis* should therefore be interpreted in a strict sense: the diagnostic spinose hypostoma must be identified from a locality before the generic name *Mucronaspis* can be applied to the specimens from the locality. Obviously this course of action creates complications, necessitating in most instances open nomenclature.

Hirnantian, and some pre-Hirnantian, dalmanitaceans referred either to *Dalmanitina* or *Mucronaspis*, and variously assigned to the species *mucronata* Brongniart, 1822, *olini* Temple, 1952, or other more recently erected ones, are almost impossible to assess, because many reported occurrences of these latest Ordovician dalmanitaceans do not illustrate hypostomata, or else the material is more or less severely distorted. A critical look at associated hypostomata is needed to prove or disprove polyphyly in these dalmanitaceans, confirm generic assignments and thus tabulate occurrences, before these trilobites are used for unequivocal dating of the latest Ordovician, as yet impossible with the data at hand. Nonetheless, *Dalmanitina* (*Songxites*) is apparently restricted to the Hirnantian.

Subfamily **DALMANITININAE** Destombes, 1972

SYNONYM. Mucronaspidinae Holloway, 1981.

DISCUSSION. Holloway (1981) distinguished the Mucronaspidinae (*Mucronaspis*, *Eodalmanitina*, *Eudolatites* (*Eudolatites*) Delo, 1935, *E.* (*Banilites*) Destombes, 1972, *E.* (*Deloites*), *Retamaspis* Hammann, 1974 and ?*Chattiaspis* Struve, 1958) from the Dalmanitininae (*Dalmanitina*, *Crozonaspis*) exclusively on thoracic and pygidial characters. Many characters listed by Holloway (1981) are couched in jargon (well rounded as against not strongly rounded pleural bands; thick and deep as against sharply impressed pleural furrows; shallow and sharply impressed as against sharply impressed interpleural furrows), while other characters differ little in each subfamily (posteriorly elongated posterior projections of thoracic pleural tips, which may be spinose as against rounded; thoracic and pygidial facets (essential to enrollment), either wholly as against essentially non-furrowed). If almost straight pygidial pleural furrows are typical of the Dalmanitininae, none of the Chinese *Dalmanitina* are correctly assigned. While pygidial doublures are said to be narrow in the Dalmanitininae, and broad in the other subfamily, this feature is still contentious at the specific level, for example in *Stenopareia linnarssoni* (Holm, 1882) (Lane 1979: 16). Of Holloway's criteria between the two subfamilies, perhaps the slope of the pleural bands is distinctive, but the same morphology is recurrent in dalmanitaceans. In any event, this last criterion alone is insufficient for subfamilial distinctness; at best, one could envisage tribal status for spinose hypostomata, but present data are insufficient for this taxonomic status.

Genus ***DALMANITINA*** Reed, 1905

TYPE SPECIES. *Phacops socialis* Barrande, 1846.

DISCUSSION. Two distinct subgenera are recognized within this genus: *D.* (*Thuringaspis*) (type *D.* (*Thuringaspis*) *osiris* Struve, 1962) (recently discussed by Cocks & Fortey 1986) and *D.* (*Songxites*) Lin, 1981, which has been accorded generic status by VandenBerg *et al.* 1984, as it was assigned to the Mucronaspidinae. Until further data from Dob's Linn (see below) are presented, subgeneric status is preferable.

Subgenus *SONGXITES* Lin, 1981

TYPE SPECIES. *Dalmanitina* (*Dalmanitina*) *wuningensis* Lin, 1974.

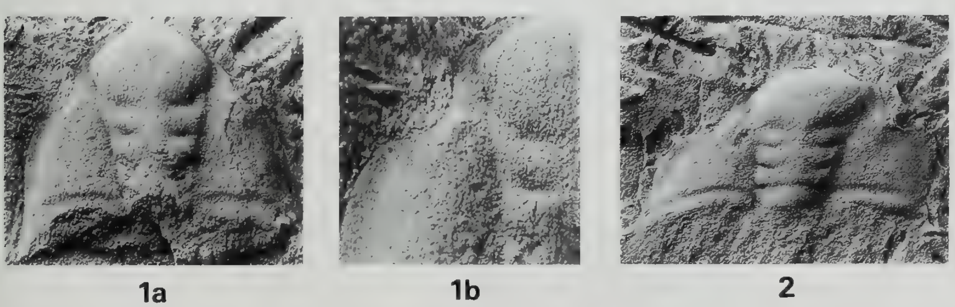
DISCUSSION. Siveter & Ingham *in* Siveter *et al.* 1980 indicated that the reduced palpebral lobe of *D.* (*Songxites*) *cellulana* of these authors was the most distinctive feature of an as yet unnamed genus, which would also encompass the Dob's Linn dalmanitacean described below. Lin's (1981) erection of the subgenus *D.* (*Songxites*) appears to have pre-empted this question as *D.* (*Songxites*) *wuningensis*, *D.* (*Songxites*) *darraweitensis* Campbell, 1973 (see VandenBerg *et al.* 1984) and *D.* (*Songxites*) *cellulana* are very closely related by the possession of reduced palpebral lobes and eye ridges in contact with the axial furrow, opposite (tr.) the 3p lobes. The hypostomata of *D.* (*Songxites*) *darraweitensis* and *D.* (*S.*) *cellulana* have approximately equal lengths and widths, significant lateral and posterior borders, but are non-spinose, as is apparently *D.* (*Songxites*) sp. (nov.) discussed below (Siveter & Ingham *in* Siveter *et al.* 1980: 201). This suggests that an assignment to *Dalmanitina* (as opposed to *Mucronaspis*) is indicated.

***Dalmanitina* (*Songxites*)** sp. (nov.)
Figs 1–2

1980 *Mucronaspis* sp. Siveter & Ingham *in* Siveter *et al.*: 200, 201.

MATERIAL. Material collected in 1979 by this writer consists of six complete cranidia (and five less complete ones), three incomplete pygidia, one fragmentary thoracic segment, and a fragmentary hypostoma. It comes from a level 10 cm below the *extraordinarius* Band at Dob's Linn, Scotland. Additional material has been alluded to, including librigenae (Siveter & Ingham *in* Siveter *et al.* 1980: 201).

DISTINCTIVE ATTRIBUTES. Maximum (tr) width of fixigenae same as maximum width (tr) of frontal glabellar lobe: fixigenae thus very wide. Lateral border furrow shallow, not reaching more incised posterior border furrow. Genal spine short and stout, approximately as long along its length as distal part of posterior border (exsag). Posterior branch of facial suture reaching border at a point (tr) from middle of 3p lobe. Anterior branch of facial suture delimiting a progressively narrower (tr) fixigena, merging into a narrow (exsag) frontal border, absent in front of central third of frontal glabellar lobe. A slightly anteromesially elongated protuberance, opposite (tr) proximal end of 3p furrows, slopes equally in all directions; in so doing, this protuberance reaches the facial suture, which is not dorsally deflected. Protuberance presumably an obsolete palpebral lobe, but librigenae or complete cephala essential to confirm this;



Figs 1–2 *Dalmanitina* (*Songxites*) sp. (nov.). Two differentially preserved inner moulds of cranidia, Fig. 2 showing obvious shearing; from a level 10 cm below the *extraordinarius* Band, Dob's Linn, Scotland. Figs 1a, 1b, BM(NH) It.20480; 1a, × 6·8; 1b, lateral view showing presumed obsolete palpebral lobe and anterior fixigenal area, × 13 (counterpart, not illustrated, BM(NH) It.20480a, shows an undamaged occipital segment). Fig. 2, BM(NH) It.20481, × 3·5.

eyes, presumably, degenerate. 2p furrows transverse, proximal end of 1p furrows slightly posteriorly directed, central part of occipital furrow shallower than distal parts.

Posterior part of hypostoma not preserved, with a distinct lateral border. Pygidial pleural furrows twice as deep and twice as wide as interpleural furrows, anteriormost four pairs evenly curved posterolaterally.

All the material consists of inner and outer moulds; exoskeleton probably very thin and unornamented.

DISCUSSION. The presumed obsolete palpebral lobe, the absence of an eye-ridge (as previously noted by Siveter & Ingham *in* Siveter *et al.* 1980: 205), and a significant anterior fixigenal area are the unique characters of this species, which should be named when the extant material is brought together.

Genus ***MUCRONASPIS*** Destombes, 1963

TYPE SPECIES. *Dalmanitina* (*Mucronaspis*) *termieri* Destombes, 1963.

Mucronaspis danai (Meek & Worthen, 1866)

Figs 3–9

1866 *Dalmania Danæ* Meek & Worthen: 264.
1868 *Dalmanites Danæ* (Meek & Worthen) Meek & Worthen: 363; pl. 6, figs 1a–f.
1917 *Dalmanites danai* (Meek & Worthen) Savage: 147; pl. 8, figs 16, 17.
1940 *Dalmanites danae* (Meek & Worthen); Delo: 40; pl. 3, figs 24, 25.

TYPES. Meek & Worthen's (1868) first illustrations of the species, along with the original description marginally modified, were based on four distinct specimens: a cephalon, a pygidium, an hypostoma, and an incomplete outstretched individual, with a major part of the left side wanting. Two institutions now hold A. H. Worthen's types. The University of Illinois at Urbana-Champaign (UI), under lot X-98 (and 11635), has (a) a complete individual, with the posterior half of the thorax wanting (this specimen has never been illustrated and is not a type), (b) a pygidium (illustrated in Delo 1940: pl. 3, fig. 25; not the original of Meek & Worthen 1868: pl. 6, figs 1d, 1e), and (c) a cephalon claimed to be a syntype of *M. danai* (original of pl. 6, figs 1b, 1c of Meek & Worthen 1868; Hansman & Scott 1967), reillustrated in Delo (1940: pl. 3, fig. 24), but this writer has been unable to examine this specimen recently. Delo (1940) referred to the complete individual above as the holotype, and the pygidium as a paratype (in the text), but in the plate explanations the pygidium and the cephalon are treated as paratypes. This is not, however, considered a designation of a lectotype (which would be invalid in any event).

The Worthen collection in the Illinois State Geological Survey, formerly Illinois State Museum [ISGS(ISM)], holds a syntypic lot of five specimens (Kent 1982): (a) a complete specimen, with much of the left side wanting (original of Meek & Worthen 1868: pl. 6, fig. 1a; 2184-1); (b) a teratological pygidium, with the right pleuron damaged, never illustrated or referred to (2184-2); (c) a cephalon, with most of the right gena missing, never illustrated or referred to (2184-3); (d) a cranidium, with most of the occipital segment broken off, never illustrated or referred to (2184-4); and (e) a pygidium, very probably the original of Meek & Worthen: pl. 6, figs 1d, 1e (2184-5). No hypostoma is thus present in these type collections; two specimens can be identified as syntypes (ISGS 2184-1 and 2184-5), in addition, apparently, to the cephalon in UI X-98. Meek & Worthen's (1866, 1868) measurements refer only to the complete individual (although mention is made in the discussion of an enormous pygidium five inches in length). ISGS 2184-1 is herein designated **lectotype** of *Dalmania danae* (recte *danai*) Meek & Worthen 1866; ISGS 2184-5 becomes a paralectotype, as apparently does the cephalon in UI X-98. The syntypic hypostoma appears lost, which is not surprising in view of the adventures of the Worthen collections (Kent 1982). From the preceding, it is clear that this writer accepts as syntypes only those specimens illustrated or referred to in the original description of the species; it is possible that some of the specimens referred to above, but not considered paralectotypes, were indeed syntypes. Formal indication that they were used by

Meek & Worthen (1866) must be presented, however, before they are added to the paralectotype list.

Savage's (1917) drawings of hypotypes (lot UI X-910, topotype cephalon and pygidium) are imprecise, the pygidium particularly so (notably the posterior part of the axis); the upturned posterior spine can, however, be observed on the original.

Mucronaspis danai is commonly cited as being erected in 1865, but Hansman & Scott (1967) have shown that the December issue of the Proceedings of the Philadelphia Academy of Natural Sciences was published in 1866. Savage's (1917) publication was also published as an extract in November 1913 (Notice between pp. 66 & 67, Savage 1917), with a different pagination.

OCCURRENCE. The syntypes are from an erosional channel of the basal Leemon Formation, along the east bank of the Mississippi river, 5900 ft (1·8 km) NNW of the railroad track and road intersection on the eastern edge of Thebes, Alexander county, Illinois. J. H. Stitt has collected this species from the Late Ordovician Edgewood Group (probably from the Cyrene Member), from a stream outcrop immediately south of 'Ebenezer Church' (Elsberry 15 minute quadrangle, 1934 edition), Lincoln county, 18 mi (29 km) southeast of Louisiana, Missouri.

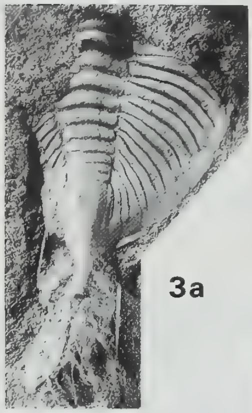
ESSENTIAL ATTRIBUTES. Maximum width of glabella anteriorly, slightly posteriorly of junction of axial and lateral border furrows, 46% of width measured across (tr) occipital segment. 2p furrows essentially transverse, but arched anteriorly, 1p furrows faintly and, more commonly, distinctly posteriorly directed proximally, distal 1p lobes isolated by shallow inner (exsag) furrows, more incised on smaller specimens. Frontal glabellar lobe with auxiliary impression patterns, median posterior impression well developed, stellate, with apparently six rays. Palpebral lobe forms highest part of cephalon; eyes with 37 (or 36?) dorsoventral files, commonly with 10 lenses per file (for a total of approximately 300 lenses), but with as few as 8 lenses per file in smaller specimens. Posterior branch of facial suture reaches marginal furrow at a point across (tr) from 1p furrow, then turns sharply posteriorly across convex border and reaches margin at a point across (tr) occipital furrow. Posterior border furrow deeply incised, meeting marginal furrow, which is the junction of differently dipping border and inner parts of genae (and thus not incised). Frontal border narrow, commonly more or less crushed. On a well preserved topotype specimen, 32 mm long (sag), frontal border consists of an inner portion 0·5 mm long (sag & exsag), separated from an outer portion (librigenae) by the dorsal suture; outer portion ranging from a feather edge (sag) to 2 mm (exsag) anterolaterally of the frontal glabellar lobe. Genal spines half as long as sagittal length of cephalon.

Hypostoma subquadrate with six marginal denticles, incipient on a small individual. Border somewhat convex, significantly longer posteriorly than laterally, set off by distinct furrows. Thoracic segments deeply furrowed, with a stout posteriorly directed distal spine.

Pygidium with 8 (and an incipient ninth) deeply furrowed pleurae, posterior one exsagittal, posterior bands sloping more steeply to interpleural furrow than anterior bands. Pleural and interpleural furrows not reaching margin, former slightly more incised (longer exsag), anterior bands slightly longer than posterior bands across border. Axis with 11 distinct axial rings and a post-axial piece continuing into a posterior spine upturned at approximately 20°; length of spine (sag) same as length (sag) of anteriormost 6 or 7 axial rings, depending on the specimen. Spine and post-axial piece continuing at same height.

Ornamentation poorly known as observed only in the following instances. Pygidium probably smooth, hypostoma with scattered tubercles on anterior lobe of median body, rare to absent on posterior lobe; genae, inward of posterior and marginal furrows, covered with irregular shallow 0·8 mm depressions, lateral cephalic border with granules.

DISCUSSION. This species is almost identical to *M. mucronata* dorsally, and may eventually be synonymized with it when the relationships between *M. mucronata* and *M. olini* have been redefined. *M. danai* differs from *M. mucronata* by its tendency to have a more flaring outward (wider, tr) frontal glabellar lobe; maximum width of the glabella in *M. mucronata* is half width across occipital segment. The hypostomata, though, differ more markedly: *M. danai* has fewer tubercles and its anteriormost marginal denticles are opposite (tr) the proximal end of the median furrow while in *M. mucronata* these denticles are more posterior, nearly opposite (tr) the middle (sag) of the posterior lobe of the median body, and, furthermore, the tubercles tend to coalesce. The median furrow of the hypostoma in *M. danai* is also more incised than in *M. mucronata*.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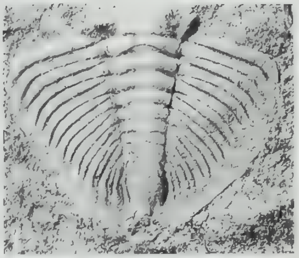
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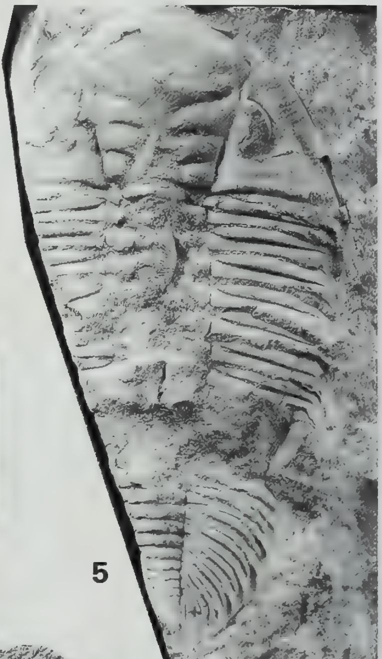
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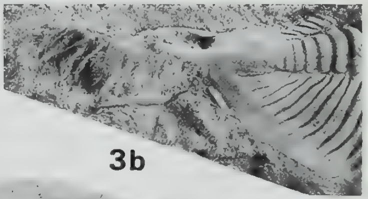
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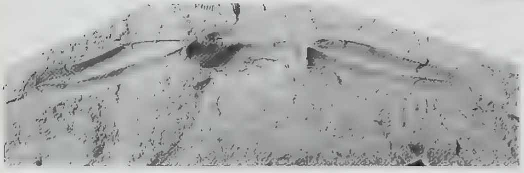
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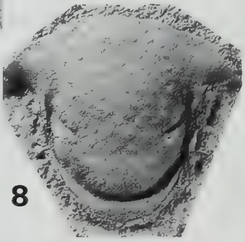
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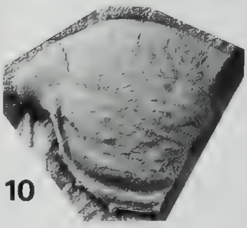
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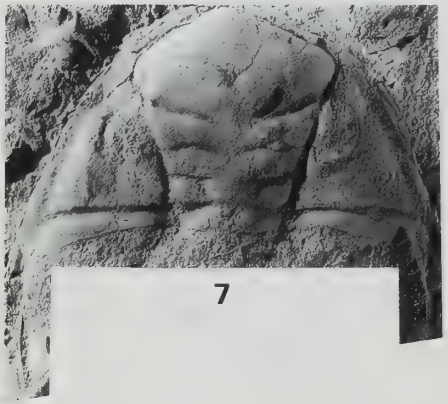
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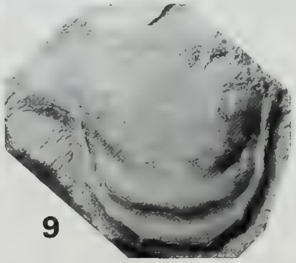
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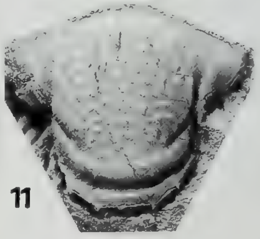
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Mucronaspis mucronata (Brongniart, 1822)
Figs 10, 11

1822 *Asaphus mucronatus* Brongniart: 24.
1822 Asaphe mucroné, *Entomostracites caudatus* de Wahlenberg; Brongniart: 144; pl. 3, fig. 9.
1952 *Dalmanitina mucronata* (Brongniart) Temple: 10; pl. 1, figs 1–3 , 5–8; pl. 2, fig. 1.
1981 *Mucronaspis mucronata* (Brongniart) Lespérance & Sheehan: 232; pl. 3, fig. 4; pl. 4, figs 1, 2, 4.
1982 *Mucronaspis mucronata mucronata* (Brongniart); Owen: 271, figs 1A, 1B.

TYPES. Lectotype cephalon and paralectotype pygidium selected by Owen (1982), Uppsala University, from the 'Dalmanitina' Beds, Västergötland, Sweden.

DISCUSSION. The above synonymy list includes only those illustrated occurrences that can obviously be referred to the species [but the Percé hypostoma included in this list (Lespérance & Sheehan 1981), and reillustrated here for comparison with *M. danai*, with another from the same locality (Figs 10, 11), could conceivably be *M. olini* (Temple 1952)].

Our understanding of this species must still be founded on Temple's (1952) careful study. He has detailed its intraspecific variability and occurrences, but did not record the spinose hypostoma. He distinguished *mucronata* from *olini* almost exclusively on pygidial characteristics, and in fact Lespérance & Sheehan (1981) could not distinguish cephala of the two species, although this distinction is obvious using the pygidia. Because of this, this writer remains convinced that careful bed by bed collecting may eventually prove or disprove suggestions that *olini* is only a geographical variant (or ecologically controlled) subspecies of *mucronata*, and thus the two species should be kept separate until conclusively proven otherwise.

A complete hypostoma of *M. mucronata kiaeri* (Troedsson, 1918) (Owen 1982, from the Rawtheyan and Hirnantian of the Oslo region, Norway) is unknown, but at least 'a small spine base a short distance out from the sagittal line' is known (Owen 1982: 274), indicating that *kiaeri* is assigned to the proper genus.

Family **TRINUCLEIDAE** Hawle & Corda, 1847

Trinucleid trilobites occur within the Hirnantian, but they are very uncommon. *Cryptolithus portageensis* sp. nov., described below, occurs in the Percé area. A trinucleid brim fragment has been reported between extensive Hirnantian brachiopod and trilobite faunas and below the *persculptus* Zone at Keisley, northern England (within unit 9 of Wright 1985: 267). Perhaps more significantly, a fragment of a tretaspid (suggesting the *Tretaspis seticornis* (Hisinger, 1840) group) occurs in northern Wales (in the type region of the Hirnantian) within a brachiopod-

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Figs 3–9 *Mucronaspis danai* (Meek & Worthen, 1866). Figs 3–5, 7, and 9 types and topotypes from north of Thebes, Illinois, Leemon Formation (formerly referred to the Edgewood Group); Figs 6 and 8, from stream outcrop near 'Ebenezer Church' (longitude 90° 53′ 19″, latitude 39° 12′ 57″), northeastern Missouri, Edgewood Group (Late Ordovician). Fig. 3, pygidium, latex cast of outer mould with exoskeleton showing upturned spine, posterior part preserved on original; 3a UMC 16590a, × 1 (outer mould UMC 16590, not illustrated); 3b, lateral view emphasizing spine, × 1. Fig. 4, inner mould, paralectotype pygidium, ISGS 2184-5, × 1·2. Fig. 5, inner mould, incomplete individual, **lectotype** (herein selected), ISGS 2184-1, × 0·7. Fig. 6, thoracic segment, outer mould with exoskeleton, stout spine on pleural tips can be discerned, UMC 16591, × 1·2. Fig. 7, cephalon, inner mould, UMC 16592, × 0·9 (partial outer mould with exoskeleton shows a complete eye, UMC 16592a, not illustrated). Fig. 8, inner mould, small hypostoma with incipient denticles, UMC 16593, × 3·9. Fig. 9, inner mould, incomplete hypostoma with six denticles, UMC 16594, × 1·4. [ISGS: Illinois State Geological Survey, Champaign, Illinois; UMC: University of Missouri at Columbia, Columbia, Missouri.]

Figs 10–11 *Mucronaspis mucronata* (Brongniart, 1822). Inner moulds of incomplete hypostomata, Côte de la Surprise Member, White Head Formation, 17 km west-northwest of Percé, Québec. Fig. 10, showing three denticles on left side, posteriormost one present, GSC 83013 (GSC 83013a, counterpart with exoskeleton, not illustrated), × 1·8. Fig. 11, showing a total of four denticles (posteriormost two denticles present on counterpart with exoskeleton, GSC 21909a, not illustrated), GSC 21909, × 1·9.

dominated [*Hirnantia sagittifera* (M'Coy, 1851), *Crytothyrella* sp. and *Plectothyrella platystrophoides* Temple, 1965] community at the Graig-Wen quarry, Powys (SJ 1018 0930) (J. T. Temple *in coll.* & personal communication 1985).

Genus ***CRYPTOLITHUS*** Green, 1832

TYPE SPECIES. *Cryptolithus tessellatus* Green, 1832.

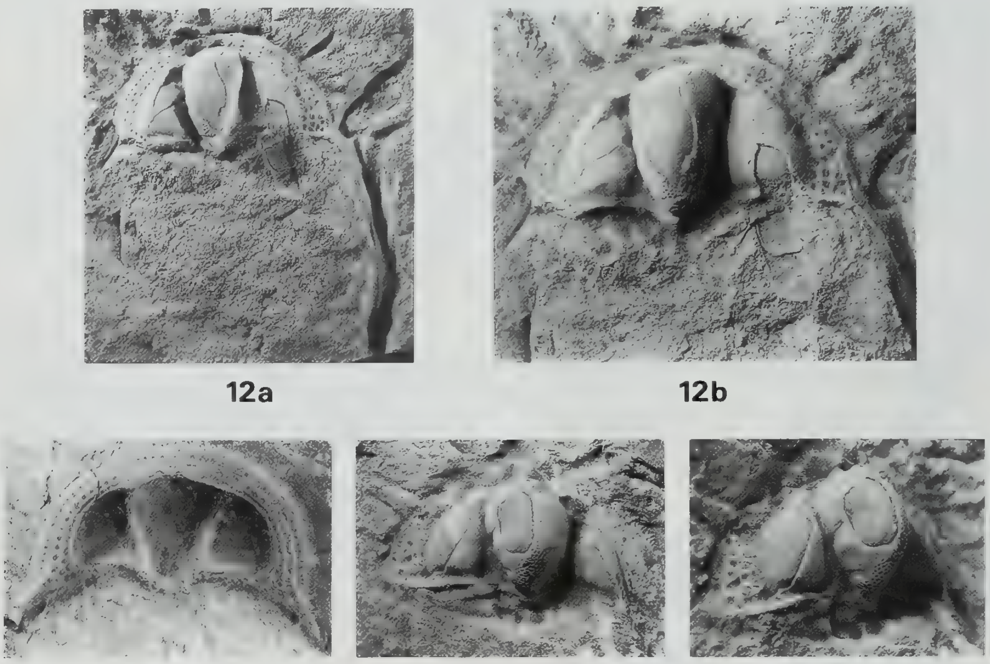
Cryptolithus portageensis sp. nov.

Figs 12–14

1974 *Cryptolithus* n. sp. Lespérance: 15.
1981 *Cryptolithus* n. sp. Lespérance & Sheehan: pl. 3, fig. 2.
1985 *Cryptolithus* n. sp. Lespérance: 845.

TYPES. Holotype: cephalon Geological Survey of Canada, Ottawa (GSC) 21914 (previously illustrated in Lespérance & Sheehan 1981), paratype cephala GSC 82988 (ventral view of lower lamella of fringe) and 82989. Also known from an additional six more or less complete cephala. From a small tributary to the Portage River, 17 km WNW of Percé, Côte de la Surprise Member, White Head Formation, Hirnantian (Lespérance 1974, and this volume, p. 242).

DIAGNOSIS. A species of the genus without glabellar furrows or pits, but with auxiliary impression patterns. The species has complete E_1, I_1 and I_n arcs, but no I_2 arc. Sagittal and immediately adjacent parts of glabella distinctly reticulated.



Figs 12–14 *Cryptolithus portageensis* sp. nov. Specimens with exoskeleton, same locality as Figs 10–11. Figs 12a, 12b, **holotype**, GSC 21914; 12a, showing length of genal spines, × 2·8; 12b, showing well girder on left side, × 4·3. Fig. 13, lower lamella of fringe, paratype GSC 82988, × 3. Figs 14a, 14b, incomplete cephalon showing ornamentation and glabellar auxiliary impression patterns, paratype GSC 82989; 14a, × 3·4; 14b, lateral view, × 3·8.

Description. Sagittal length of cephalon twice maximum width measured across posterior margin. Genal spines slender, flaring outward, then inward distally, 1·5 times length of cephalon. Sagittal tubercle on glabella, slightly in front of glabellar mid-point (excluding occipital segment). Posterior margin of occipital segment entire, not drawn out by a spine, nor possessing a tubercle. Occipital furrow and posterior margin furrow wide (sag, exsag), deep, but occipital shallower. Glabellar furrows or pits absent, but three pairs of darker, slightly impressed auxiliary impression patterns present on sides of glabella, a short distance from axial furrow. Posterior pair comma-shaped, with a more strongly curved portion ventralmost, almost touching occipital furrow, elongated essentially perpendicularly to axial furrow, approximately 1 mm in greatest dimension; second pair circular, approximately 0·6 mm in diameter; anterior pair much as posterior pair, but ventral portion not posteriorly elongated, 0·6 mm along its greatest length, situated essentially transversely to glabellar tubercle (measurements taken from paratype cephalon GSC 82989).

Prominent girder list present on upper lamella of fringe; another list, between I_1 and I_n only present on posterior half of fringe. Lower lamella of fringe with pseudo-girder between I_1 and I_n, girder continuous onto genal spine; both girder and pseudo-girder attenuated toward sagittal line. Genae smooth, central and highest part of glabella (sag, exsag) reticulated for a width of approximately 1 mm (tr) (as present on GSC 82989), but ornamentation unknown on anteriormost, and subvertical, portion of glabella.

Following the orientation suggested by Hughes *et al.* (1975: 547), frontal part of fringe horizontal, laterally gentle sloping downward. Arcs E_1, I_1 and I_n complete; I_2 absent. Half fringes with 24–25 pits in E_1 18–20 in I_1, and 18–19 in I_n arc; 8–10 smaller flange pits present posteriorly, and 6–8 occur along the posterior margin of the fringe.

Dimensions. All the type material is slightly laterally compressed; measurements are in mm.

	Length (sag)	Width across posterior margin
GSC 21914	5·6	11·7 (est.)
GSC 82988	6·3	12·9
GSC 82989	—	11·5 (est.)

Discussion. Glabellar auxiliary impression patterns are known in Caradoc species of *Cryptolithus* (Whittington 1968: pl. 87, figs 6, 10; pl. 88, fig. 11; pl. 89, fig. 1). The low number of pits, particularly the absence of an I_2 arc, as well as a different glabellar ornamentation, distinguish *C. portageensis* sp. nov. from *C. stoermeri* Owen, 1980, from the uppermost Husbergøya Formation (upper Rawtheyan) of the Oslo region. *C. portageensis* sp. nov. is nearest *C. kosoviensis* Marek, 1952 (uppermost Králův Dvůr Formation, Rawtheyan?, Bohemia), which however has a frontally incomplete I_n arc; only the posterior half of the glabella of *kosoviensis* is reticulated, as is part of the inner posterior cheeks (Přibyl & Vaněk 1969: 104). Hughes *et al.* (1975) have questioned the assignment of *kosoviensis* to *Cryptolithus*, but the similarity of *portageensis* to *kosoviensis* suggests that the Bohemian species is correctly assigned to *Cryptolithus*.

Family **PHACOPIDAE** Hawle & Corda, 1847

Although the genus *Acernaspis* apparently first occurs with the onset of the *acuminatus* Zone, Lespérance & Letendre (1982: 329) have drawn attention to a new genus of this family that first occurs in the Belgian Ashgill.

Genus *ACERNASPIS* Campbell, 1967

Type species. *Phacops orestes* Billings, 1860.

Remarks. *Acernaspis* (subgenus?) *norvegiensis* Lespérance & Letendre, 1982 is herein reillustrated (Fig. 15) to show its distinctness from other species of the genus. It is the only known species within *Acernaspis* which has granules and pustules, many of the latter being perforated. It may be noted here that this species is associated with another species of *Acernaspis* within '6b' of the Asker region, Norway (Lespérance & Letendre 1982: 336).

Subgenus *ACERNASPIS* Campbell, 1967

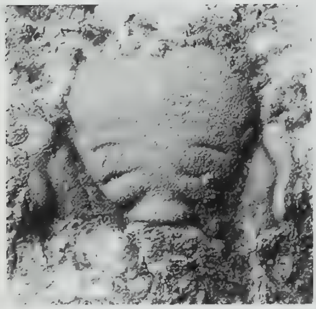
Diagnosis. Primitive phacopids with continuous vincular furrows, which may be anteriorly shallower. Ornamentation variously with punctae or smooth, but more commonly granulose (Lespérance & Letendre 1981: 199).

Remark. The use of subgenera within *Acernaspis* has been amply discussed by Lespérance & Letendre 1981, and need not be repeated here.

Acernaspis* (*Acernaspis*) *salmoensis sp. nov.
Figs 16–19

1981 *Acernaspis* sp. Lespérance & Letendre: 197.
1982 *Acernaspis* sp. Lespérance & Letendre: 329.
1982 *Acernaspis* (*Acernaspis*) n. sp.? Lespérance & Letendre: 332; pl. 1, fig. 16.
1985 *Acernaspis* n. sp. Lespérance: 845.

Types. Holotype: GSC 69146, previously illustrated (Lespérance & Letendre 1982). Paratypes: GSC 82990, incomplete cranidium; GSC 82991, a pygidium; and GSC 82992, incomplete cephalic doublure.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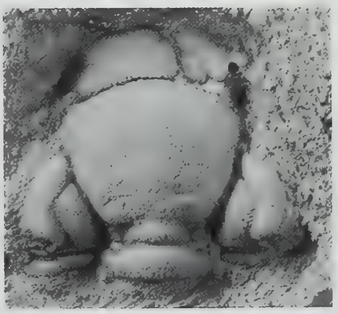
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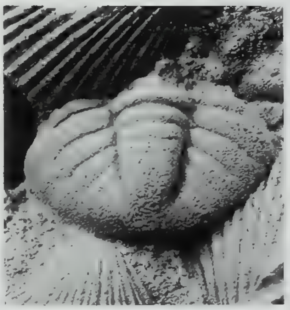
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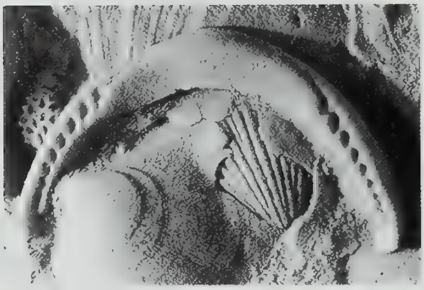
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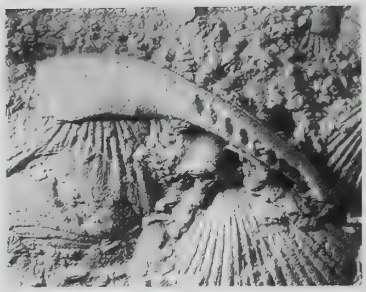
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Fig. 15 *Acernaspis* (subgenus?) *norvegiensis* Lespérance & Letendre, 1982. Incomplete cranidium with exoskeleton, upper half of '6bα' (Solvik Formation: Worsley 1982: 165), Spirodden peninsula, Asker region, Norway; PMO 106·509, × 9·5. [PMO: Paleontologisk Museum, Oslo.]

Figs 16–19 *Acernaspis* (*Acernaspis*) *salmoensis* sp. nov. Specimens with exoskeleton, Becscie Formation, Anticosti Island, Québec. Fig. 16, incomplete cranidium, paratype GSC 82990, × 7·1. Fig. 17, pygidium, paratype GSC 82991, × 6·8. Fig. 18, cephalic doublure showing vincular furrow, **holotype** GSC 69146, × 7·3. Fig. 19, incomplete cephalic doublure, paratype GSC 82992, × 5·3. [GSC: Geological Survey of Canada, Ottawa.]

OCCURRENCE AND MATERIAL. Only known from the Rhuddanian Becscie Formation of eastern Anticosti island, Québec. Paratypes from roadside outcrop on northern side of road parallel to, and south of, Salmon River, from a level 4 m above lowermost occurrence of the species. This outcrop extends westward from a stream emptying into the river, and is 960 m west of longitude 62° 18′ 00″ and 250 m south of latitude 49° 24′ 00″. This level has yielded approximately 45% of the known material of the species, and the level 4 m below it another 45%. This lowermost level is 45 m above the base of the Becscie Formation (Lespérance 1985: 845). The species also occurs at the 'major falls' along the Salmon River, at 'pool 16' (9·5 km west of the previous locality), and the holotype is from an outcrop along the road leading to Baie de la Tour, 0·8 km north of the main road (approximately 27 km to the northwest of the paratypes; see also Lespérance & Letendre 1982: 334). Extant material of the species includes approximately 10 cephalic doublures, 35 cranidia, 60 pygidia and a few incomplete thoracic segments and librigenae.

DIAGNOSIS. A species of *Acernaspis* (*Acernaspis*) with a very shallow anterior vincular furrow and a posterior vincular furrow with dividing walls between fossulae; dorsal sutures functional and ornamentation consisting of microgranules.

DESCRIPTION. Glabella expanding forward, widest across frontal glabellar lobe, with a width ratio of 8:5 with width (tr) of occipital segment. 3p furrows bicomposite, distal part impressed, proximal part faintly, as 2p furrows. Distal 1p lobe isolated, below level of 2p lobe and distal part of occipital segment. 1p furrow continuous, poorly incised and shallow sagittally. Occipital furrow incised, continuous. Palpebral furrow incised, extending from axial furrow anteriorly to a point transverse from occipital furrow. Posterior border furrow wide (exag), incised. Palpebral lobes below level of central part of glabella, convex and thus bent downward distally. Dorsal sutures functional. Eyes with a minimum of 14 dorsoventral files, with 3–5 lenses per file.

Anterior part of vincular furrow marginal and ventral, as anterior and anterolateral part of subvertical doublure slopes very steeply posteriorly. Anterolateral section of anterior part of vincular furrow broadly incised, but sagittally barely perceptible and very shallow. Posterior part of vincular furrow scalloped, with 8 or 9 fossulae, with dividing walls between fossulae reaching approximately the mid-point between the bottom of the fossulae and the bounding walls. Anterior half of proximal bounding wall of posterior vincular furrow vertically below adjacent/distal wall, while posterior half of proximal bounding wall of posterior vincular furrow vertically shorter than outer, adjacent distal wall.

Pygidium wider than long (as 8:5), axis with 7 axial rings, not reaching posterior margin. Axial ring furrows transverse, progressively shallower posteriorly. Pleurae with 4 pygidial ribs, very faintly furrowed; distal third of pleural fields unfurrowed. Articulating half-ring cut in middle by facet; furrow between this half-ring and anteriormost rib apparently continuous to margin.

Ornamentation consisting of microgranules (densely packed 0·01–0·04 mm granules, better developed on cephalic doublure, including the anterior part of the vincular furrow), probably modified by surficial weathering.

DIMENSIONS. All lengths given are sagittal and all widths are transverse; measurements are in mm.

	GSC 69146	GSC 82992
Width of cephalon	7·3	—
Length of cephalic doublure	1·16	1·91

Paratype pygidium (GSC 82991) has a width of 5·0; its total length is 3·1, which includes a length of 0·20 for the articulating half-ring; length of axis, including articulating half-ring, 2·6. Paratype cranidium (GSC 82990) has a length of 3·8, and widths of 2·2 for the occipital segment and 0·6 for the palpebral lobe.

DISCUSSION. The very shallow anterior part of the vincular furrow sets this species apart from all others within the subgenus. The taxon closest to it appears to be *Acernaspis* (*Murphycops*) *skidmorei* (Lespérance, 1968) (Lespérance & Letendre 1981), which has no anterior vincular furrow and in which the anteriormost part of the cephalic doublure is vertical. *Acernaspis* (*A.*) *salmoensis* sp. nov., in this regard, appears as an ideal ancestor for *A.* (*Murphycops*) *skidmorei*, of lower Idwian age. The lowest *Acernaspis* sp. from the Becscie Formation of western Anticosti, near Cap à l'Ours (Lespérance 1985: 845), is too poorly preserved for specific assignment.

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Note added in page proof. Additional topotype material of *Cryptolithus portageensis* sp. nov., previously not examined and from a different field collection number, contains three partial and a complete cephalon, as well as a pygidium with a damaged axis. Ornamentation on the central part of the glabella continues on the subvertical frontal lobe, but does not reach the fringe. The pygidium has a width to length ratio of 4 : 1, three interpleural furrows not quite reaching the steeply inclined border, and a fourth incipient and posterior one.

Environmental changes close to the Ordovician–Silurian boundary

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Synopsis

Most late Ordovician to early Silurian sequences show evidence of a regressive phase followed by transgression, reflecting glacio-eustatic sea-level changes. Continental glacial deposits are particularly well known from Saharan Africa, and glaciomarine deposits from Iberia and Normandy. Rapid growth of the ice caps at the beginning of the Hirnantian is reflected on clastic marine shelves by a change from mudstones to a variety of shallow marine sand facies. Withdrawal of the sea to the edges of shelves fed sand into basins to form submarine fans. Shallow carbonate shelves generally became exposed during the Hirnantian, and karstic surfaces developed. A sea-level fall of between 50 and 100 m is envisaged. The regressive deposits are usually abruptly overlain by deeper-water deposits formed during a rapid transgression. Graptolitic shales are widely developed on clastic shelves, but there is a return to shallow marine limestones on carbonate shelves. There is local evidence of oscillations of sea-level within the main Hirnantian glacial event, but it is uncertain whether these changes were eustatically controlled. It is suggested that the climate during the Hirnantian remained cold in peri-polar regions, but may have been variable in mid-latitudes and was tropical in equatorial regions. There is some palaeomagnetic evidence to suggest that continents were moving unusually fast during late Ordovician times, which might have had an influence on the growth and decay of late Ordovician ice caps.

Introduction

Most late Ordovician to early Silurian sequences show evidence of a regressive phase followed by transgression. The regressive–transgressive interval is of the same age on plates which were separate in the Lower Palaeozoic (Berry & Boucot 1973) and so satisfies the criteria for identifying eustatic sea-level changes (Fortey 1984). The fall in sea-level started at the beginning of the Hirnantian and the subsequent rise of sea-level had been largely completed before the end of Hirnantian times. A major ice cap was present on the Gondwana plate at this time and it is likely that the sea-level changes were related to the growth and decay of that ice cap.

The Ordovician–Silurian boundary, as it is now placed at the base of the *P. acuminatus* Zone, post-dates the late-Ordovician sea-level changes and falls within a period of environmental stability. Thus the often striking facies changes in the Hirnantian, and particularly the change from shallow to deeper water facies at the top of the Hirnantian, help to identify horizons immediately below the boundary between the systems, but not the boundary itself.

Duration of the eustatic changes

Different ways of estimating the duration of Hirnantian environmental changes can be made, and these produce somewhat different results. Estimates of the duration of the Hirnantian made by dividing the duration of the Ashgill, based on radiometric age determinations, by the number of stages (four) give 1·8 to 2·5 my. If the duration of the Ashgill is divided by the number of zones in the type area (eight) (Ingham 1966) the duration of the Hirnantian, which has only one zone, is 1 to 1·25 my. A value between 1 and 2 million years is probable, but more radiometric dates close to the Ordovician–Silurian boundary are needed to give more accurate estimates.

Changes in sedimentary environments

Continental glaciation. The deposits of continental ice sheets of upper Ordovician age in Saharan Africa are well known through the descriptions of Beuf *et al.* 1971, Rognon *et al.* 1972,

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and others. They recognized nearly all the features characteristic of land-based ice deposition, including glaciated pavements, striated pebbles, tillites, varved sediments and dropstones, and a wide variety of fluvio-glacial sediments (Fig. 2, section 1), some of which are associated with long esker-like ridges. Similar deposits have been recognized in South Africa (Rust 1982), and glacial deposits believed to be of a similar age have been described from west Africa, South America (see Spjeldnaes 1981 and references therein) and Saudi Arabia (McClure 1978). The late Ordovician Gondwana glaciation was clearly of continental dimensions and appears to have extended from the south pole through at least 40° of latitude. There is no evidence of a contemporary ice cap in the Ordovician northern hemisphere, which, according to palaeogeographic reconstructions, had no continental areas near the pole at that time.

Glaciomarine environments. Tilloids of glaciomarine origin were initially identified by Dangeard & Doré (1971) in Normandy, and by Hempel & Weise (1967) in Thuringia. Subsequently, glaciomarine sediments, usually consisting of pebbly mudstones, have been recognized in Brittany (Hamoumi *et al.* 1980), Celtiberia (Carls 1975), west central Spain (Robardet 1981) and Portugal (Romano & Diggens 1973–74; Young 1985).

Most of the clasts in the tilloids can be matched with carbonate or coarse clastic horizons in the underlying succession, indicating that at times the ice was grounded and caused erosion. Striated clasts are recorded from Normandy (Dangeard & Doré 1971) and Navatrasierra, western Spain (personal observation). Deposition, however, appears to have been from floating ice, as indicated by the delicately laminated nature of some of the sediments, the presence of dropstones in Brittany (Hamoumi 1981), but above all by the nature of the predominantly massive sandy mudstones which lack associated sand deposits of fluvioglacial origin. In Spain and Portugal there is evidence of regression and emergence prior to the deposition of the tilloids (Fig. 2, sections 2 and 3), and there are variable proportions of normal marine sediments interbedded with the glaciomarine sediments.

At the time of the maximum continental glaciation of the Gondwana plate, the adjacent Armorican plate apparently lacked a continental ice sheet. Here, ice was locally grounded on recently exposed shelf sediments but at times of slightly higher sea-level there was widespread floating ice from which was deposited the mainly structureless sandy mud with its dispersed clasts.

Clastic shelves. Many sequences which formed on clastic shelves show an upward passage from mudstones to shallow marine sandstones. On shelves where there was an adequate supply of sand complete upward-coarsening regressive sequences were formed starting with shelf muds and passing gradationally upwards through various shoreface facies (Fig. 1, section 1), or sometimes more abruptly into a variety of shallow marine facies (Fig. 1, sections 2 and 3). At other places where there was channelling of the shelf, massive or cross-stratified sandstones lie with a sharp erosional base on the underlying sediments (Fig. 1, section 4, might represent such a situation). When a clastic shelf or relatively shallow basin was relatively starved of sediment the regressive sequence is condensed, sometimes to as little as a metre, and may be partly calcareous, as in Västergötland (Fig. 1, section 5) where there is a thin oolite bed, or in the Yangtze Basin where a thin bioclastic limestone caps graptolite shales (Fig. 1, section 6).

At most places shallow marine sediments of the regressive phase are succeeded abruptly by mudstones with a benthic fauna indicating a deep shelf environment, or by graptolitic shales. Facies formed during the rise of sea-level are usually less than a metre thick, suggesting that the transgression was rapid.

Clastic basins. There is evidence from the Welsh Basin that the end Ordovician regression caused sediments to be carried across the marginal shelves and produced an influx of coarse clastics into previously mainly argillaceous basin environments. Pebbly mudstones of mass flow origin, thick-to-thin bedded turbidites, some of which are channelled, and some slumped units suggest the presence of substantial base-of-slope fans (Fig. 1, sections 7 and 8). At the north-west margin of the basin, fan sediments with resedimented ooids and fragmented valves of a *Hirnantia* fauna overlie trilobite-bearing mudstones, suggesting that this particular fan accumulated at no great depth.

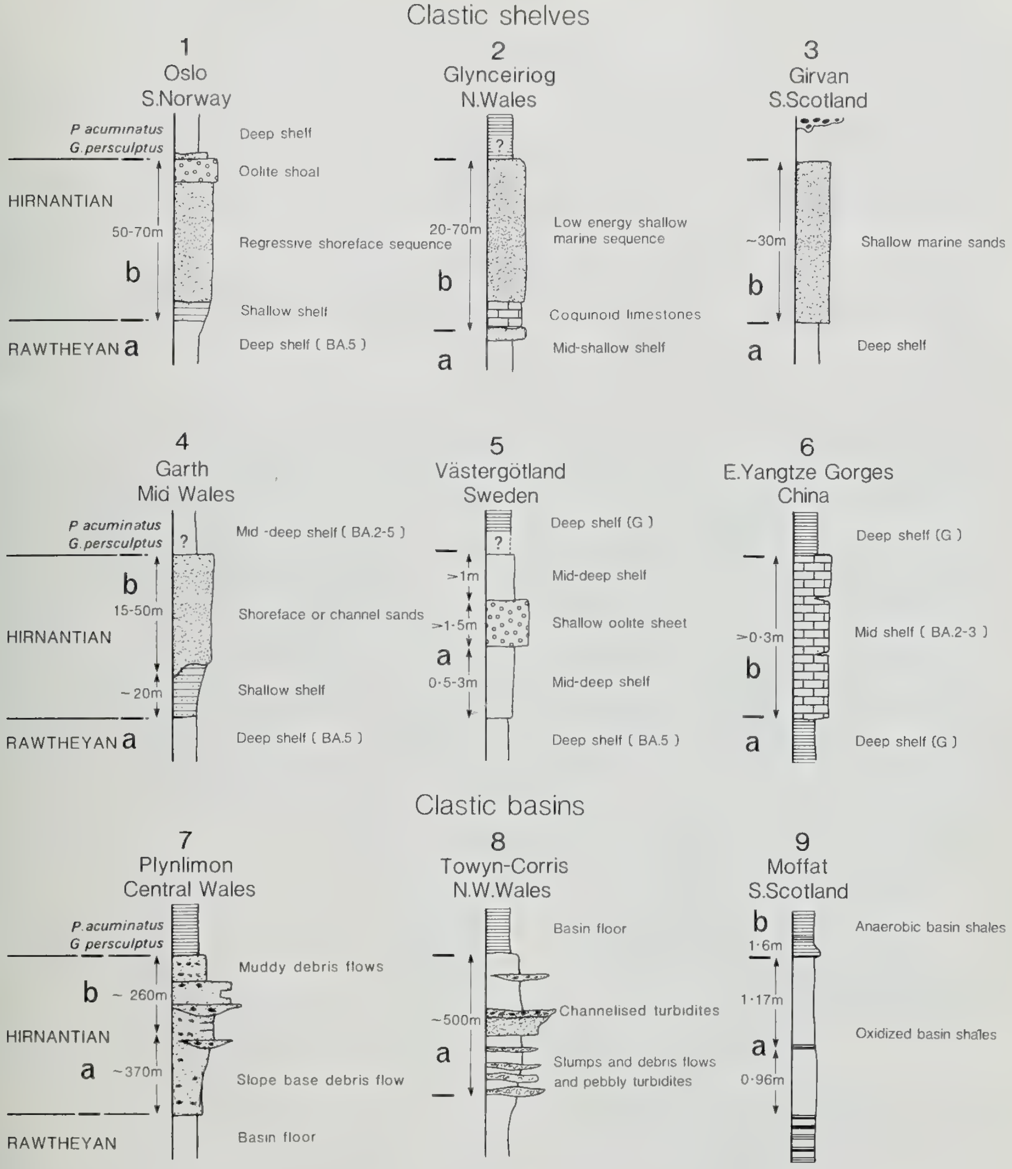
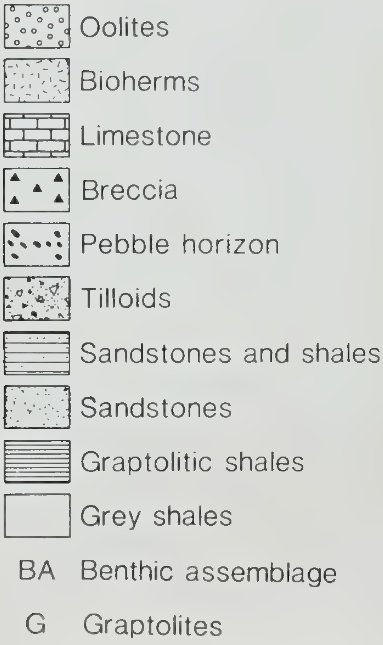

Clastic shelves
1
Oslo
S.Norway
P acuminatus
G.persculptus
HIRNANTIAN
50-70m
b
RAWTHEYAN a
Deep shelf
Oolite shoal
Regressive shoreface sequence
Shallow shelf
Deep shelf (BA.5)
2
Glynceiriog
N.Wales
?
20-70m
b
a
Low energy shallow marine sequence
Coquinoid limestones
Mid-shallow shelf
3
Girvan
S.Scotland
~30m
b
a
Shallow marine sands
Deep shelf
4
Garth
Mid Wales
P acuminatus
G.persculptus
?
b
15-50m
HIRNANTIAN
~20m
RAWTHEYAN a
Mid -deep shelf (BA.2-5)
Shoreface or channel sands
Shallow shelf
Deep shelf (BA.5)
5
Västergötland
Sweden
?
>1m
>1·5m
a
0·5-3m
Deep shelf (G)
Mid-deep shelf
Shallow oolite sheet
Mid-deep shelf
Deep shelf (BA.5)
6
E.Yangtze Gorges
China
>0·3m
b
a
Deep shelf (G)
Mid shelf (BA.2-3)
Deep shelf (G)
Clastic basins
7
Plynlimon
Central Wales
P.acuminatus
G persculptus
b ~260m
HIRNANTIAN
a ~370m
RAWTHEYAN
Muddy debris flows
Slope base debris flow
Basin floor
8
Towyn-Corris
N.W.Wales
~500m
a
Basin floor
Channelised turbidites
Slumps and debris flows and pebbly turbidites
9
Moffat
S.Scotland
b
1·6m
1·17m
a
0·96m
Anaerobic basin shales
Oxidized basin shales

Fig. 1 Generalized sections to show the sequence of environmental changes near the Ordovician/Silurian boundary. Data for the interpretations are to be found in the following references. Section 1: (a) Husbergøya Shale, (b) Langøyene Sandstone; Brenchley & Newall 1980. 2: (a) Dolhir Formation, (b) Glyn Formation; Hiller 1981; Brenchley & Cullen 1984. 3: (a) Drummuck Group, (b) High Mains Formation; Harper 1981. 4: (a) Wenallt Formation, (b) Cwm Clŷd Formation; Williams & Wright 1981. 5: (a) Dalmanitina Beds; Stridsberg 1980. 6: (a) Wufeng Formation, (b) Guanyinqiao Formation; Geng Liang-yu 1982. 7: (a) Nant-y-Moch Formation, (b) Drosgol Formation; James 1971; Cave 1979; James 1983. 8: Garnedd-Wen Formation; James 1972; James 1985. 9: (a) Upper Hartfell Shale Formation, (b) Birkhill Shale Formation; Williams 1983.

In some basins which were isolated from a source of coarse clastics there were no obvious changes in pelagic sedimentation, as in some of the graptolitic shale sequences in the Yukon (Lenz 1982; Lenz & McCracken 1982). In a rather similar graptolitic shale sequence at Dob's Linn in the Southern Uplands of Scotland, the end Ordovician regression cannot be identified but the transgression is reflected in a change from grey mudstones, without graptolites, to black graptolitic shales (Fig. 1, section 9). This change from oxidized to anoxic sediments might reflect the change from the vigorous bottom circulation of the glacial period to the more sluggish circulation following the melting of the ice caps.

The graptolitic shales, which commonly succeed the coarser clastics formed during the regression in basin environments, may contain a *G. persculptus* fauna, but may in other instances have *P. acuminatus* or even younger faunas in the lowest horizons. The local absence of the lowest Silurian graptolite zones is probably the result of erosion or non-deposition. Similar hiatuses are being increasingly recognized in DSDP cores in areas of pelagic sedimentation (Moore *et al.* 1978). For example, widespread deep-sea erosion in the Miocene is associated with periodic cold-climate events, lower eustatic sea-level and an intensification of bottom circulation (Keller & Barron 1983).

Carbonate shelves. Most of the very extensive carbonate platforms in North America and Arctic Canada appear to have been exposed at the end of the Ordovician, producing regional disconformities (Lenz 1976, 1982). The sedimentological effects of the regressive–transgressive cycle are commonly not easily recognized in shallow marine carbonate sequences. Nevertheless a late Ordovician, generally regressive, sequence culminating in a widespread oncoid bed has been recognized in Anticosti Island (Petryk 1981*a*), and this is succeeded by generally transgressive sediments (Fig. 2, section 5). At Manitoulin Island, Ontario, two karstic horizons separated by 15 cm of sediment occur close to the Ordovician–Silurian boundary in a sequence of shallow marine carbonate facies (Fig. 2, section 4). The effects of the end-Ordovician regression can also be recognized in the more offshore facies associated with carbonate mud mounds. In two of the carbonate mounds of the Boda Limestone (central Sweden) there is evidence of emergence of the mound crests, with karst surface on one mound (Fig. 2, section 6), and dripstone calcite lining fissures in the other. Graptolitic shales, formed after the transgression, mantle the



Key to Figs 1–2.

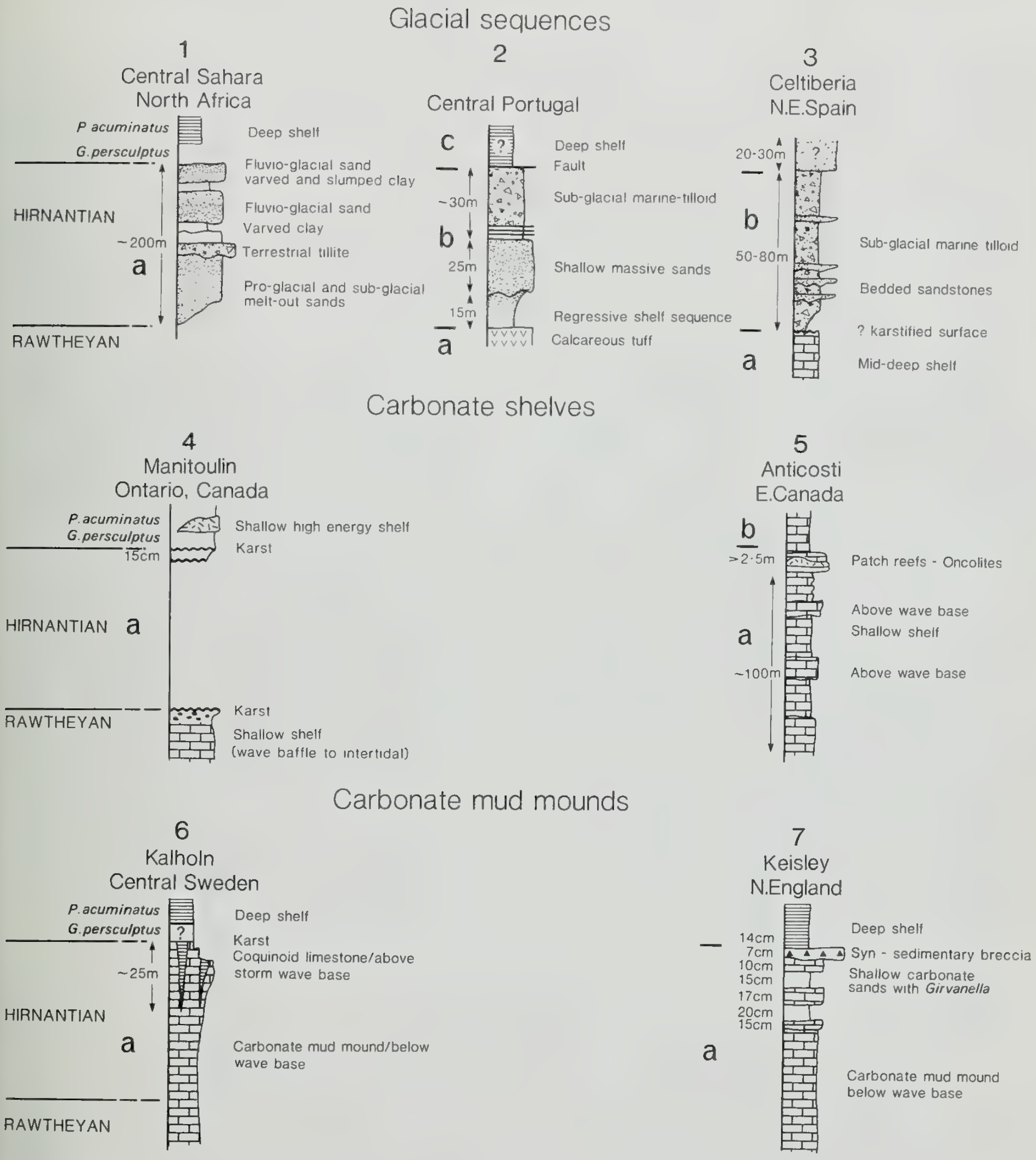

Glacial sequences
1
Central Sahara
North Africa
P acuminatus
G. persculptus
Deep shelf
Fluvio-glacial sand
varved and slumped clay
Fluvio-glacial sand
Varved clay
Terrestrial tillite
Pro-glacial and sub-glacial
melt-out sands
HIRNANTIAN
~200m
a
RAWTHEYAN
2
Central Portugal
c
?
Deep shelf
Fault
~30m
Sub-glacial marine-tilloid
b
25m
Shallow massive sands
15m
Regressive shelf sequence
a
Calcareous tuff
3
Celtiberia
N.E.Spain
20-30m
?
b
50-80m
Sub-glacial marine tilloid
Bedded sandstones
? karstified surface
a
Mid-deep shelf
Carbonate shelves
4
Manitoulin
Ontario, Canada
P. acuminatus
G. persculptus
15cm
Shallow high energy shelf
Karst
HIRNANTIAN a
RAWTHEYAN
Karst
Shallow shelf
(wave baffle to intertidal)
5
Anticosti
E.Canada
b
>2·5m
Patch reefs - Oncolites
a
Above wave base
Shallow shelf
~100m
Above wave base
Carbonate mud mounds
6
Kalholn
Central Sweden
P. acuminatus
G. persculptus
?
Deep shelf
Karst
Coquinoid limestone/above
storm wave base
~25m
HIRNANTIAN
a
Carbonate mud mound/below
wave base
RAWTHEYAN
7
Keisley
N.England
14cm
7cm
10cm
15cm
17cm
20cm
15cm
Deep shelf
Syn - sedimentary breccia
Shallow carbonate
sands with *Girvanella*
a
Carbonate mud mound
below wave base

Fig. 2 Generalized sections to show the sequence of environmental changes near the Ordovician–Silurian boundary. Data for the interpretations are to be found in the following references. Section 1: 'Unit IV'; Beuf *et al.* 1971. 2: (a) Porto de Santa Anna Formation, (b) Ribeira do Bracal Formation, (c) Ribeira Cimeria Formation; Young 1985. 3: (a) Cystoid Limestone, (b) Orea Shale; Carls 1975. 4: Georgian Bay Formation, (b) Manitoulin Formation; Copper 1978; Kobluk 1984. 5: (a) Ellis Bay Formation (up to Oncolites), (b) Becscie Formation; Petryk 1981*a*, 1981*b*. 6: Boda Limestone; Jaanusson 1979; Brenchley & Newall 1980. 7: Keisley Limestone; Wright 1985.

mounds and fill fissures in both cases. In the carbonate mound at Keisley, in northern England, the regression is reflected by the development of beds containing the alga *Girvanella* at the top of the mound. There is a final capping of breccia, a few cm thick, and this is succeeded abruptly by graptolitic shales, again marking the transgressive phase (Fig. 2, section 7).

Bathymetric changes. There is good evidence that most carbonate and clastic platforms and shelves shoaled to near sea-level or became exposed during the Hirnantian regression. Some of the platforms were already shallow before the start of the regression, but some muddy shelves which were initially below storm wave-base, suggesting water depths of several tens of metres, also became exposed (Brenchley & Newall 1980). The relief on an erosion surface below the Silurian in Iowa, USA, suggests that sea-level dropped at least 45 m (Johnson 1975). The emergence of the crests of carbonate mud mounds and the lining of fissures to a depth of nearly 30 m implies a sea-level fall of about 70 m (Brenchley & Newall 1980). A sea-level fall between 50 and 100 m seems likely though a figure of 'not more than 20 m' has been suggested by Geng Liang-yu (1982).

The widespread presence of grey mudstones with deep shelf benthic faunas prior to the regression, but graptolitic shales after the transgression, suggests that the sea-level rise might have been greater than its fall (Brenchley & Newall 1980). However, the evidence from carbonate platforms does not support this because in general early Silurian carbonates are similar to those of the late Ordovician and both suggest shallow marine environments. It may be that the development of early Silurian graptolitic facies is determined more by the preceding transgression which drowned many source areas, rather than by a substantial increase in water depths.

Although only a single regressive phase followed by transgression is apparent in many sections there is some evidence for oscillations of sea-level within the Hirnantian. Two karstic horizons representing two phases of emergence were recognized at Manitoulin Island (Kobluk 1984) and in a carbonate sequence near Oslo (Hanken 1974). Three regressive phases were identified by Petryk (1981*b*) in the upper Ordovician sequence on Anticosti Island. It is possible that these bathymetric changes might be related to phases of growth of the continental ice caps reflected by three separate horizons of till in the Saharan and South African sequences. Episodes of ice advance and retreat are now well documented in the Pleistocene record. Changes in the size of the Pleistocene ice caps produced cyclic changes in the $^{18}O/^{16}O$ isotopic record in oceanic sediments implying temperature fluctuations with a periodicity of about 20 000, 40 000 and 100 000 years (Hays *et al.* 1976) similar to those predicted by Milankovitch (1938) on astronomical grounds. A similar cyclicity might be expected in earlier glaciations, and might be represented by the three sea-level oscillations and three tills in the Hirnantian. However, the time-scale of these oscillations is still unclear.

Geochemical changes. There are very few studies of sediments close to the Ordovician–Silurian boundary which might show if the geochemistry reflected the climatic and other environmental changes. A pilot study in a relatively uniform sequence of argillaceous sediments in the type Ashgill area of northern England showed changes in carbonate, Fe and P content and in Fe_2O_3 activity at the base and/or top of the Hirnantian, which were correlated with minor changes in lithology and probably with changes in palaeobathymetry (Brenchley 1984). A study of carbon and oxygen stable isotopes in a sequence through a Boda carbonate mud mound showed changes in ^{18}O values which suggested a fall in sea-water temperature during the Hirnantian (Jux & Manze 1979). Both these studies suggest that further geochemical work might prove valuable in determining changes in sea-water chemistry and temperature during the Hirnantian.

Climatic changes. The distribution of late Ordovician glacial deposits suggests that continental ice sheets extended from the south pole through at least 40° of latitude and that there was floating ice for another 10° of latitude. The temperature of peripolar oceans would have been substantially depressed during such periods of glaciation. The effect of glaciation on the temperature of surface waters in lower latitudes is less easy to predict. Studies of surface waters at 18 000 years B.P., during the last interglacial, show marked differences between the Atlantic and Pacific Oceans, indicating there is no simple global pattern of temperature (McIntyre *et al.*

1976; Moore *et al.* 1980). Two points possibly relevant to the reconstruction of Ordovician climate do however emerge; one is that water temperatures in some tropical and temperate areas may actually be raised during a glacial episode, and the second is that notably cooler waters can develop in both temperate and tropical areas.

The widespread extension of cooler surface waters during a glaciation might explain the very broad distribution of the *Hirnantia* fauna, thought by some to be a cool-water fauna, throughout most temperate and sub-tropical regions during some part of Hirnantian times.

The possibility of elevated temperatures during a glacial phase might partly account for the apparently anomalous occurrence of Hirnantian oolitic horizons in sequences which were hitherto wholly clastic (Oslo in Norway, and Garth and Bala in north Wales). It is not necessarily a contradiction that the sequences which contain oolites also contain an *Hirnantia* fauna, since the changes of sea-surface temperatures can be substantial between glacials and interglacials, particularly in mid-latitudes.

A tentative construction of Hirnantian climate is that polar and peri-polar regions remained cool to glacial throughout the Hirnantian, mid-latitudes had very variable climatic conditions varying in time and space from cool to warm, while tropical areas in general remained hot. The climate instability and geographic contrasts of the Hirnantian were succeeded by more stable conditions in the Silurian. It is thought that the climate was in general similar to that of today, but that climatic belts were more nearly parallel to lines of latitude because of the relative absence of land in low latitudes (Ziegler *et al.* 1977).

Palaeomagnetism. The distribution of continents, based on palaeomagnetic evidence, has been reconstructed for the middle Ordovician and for the early Silurian (Ziegler *et al.* 1977; Ziegler & Scotese 1979; Scotese *et al.* 1979). Unfortunately there are no maps of comparable detail for the Upper Ordovician. Early Silurian reconstructions show Gondwanaland lying in high southern latitudes and other continents spread across the southern hemisphere and into mid-northern latitudes. No continents are located in high northern latitudes.

There is some evidence from the shape of the apparent polar-wandering paths of the Ordovician that the continents must have moved unusually fast in late Ordovician times, to create the Lower Silurian palaeogeography. Some confirmation of this rapid movement comes from a wealth of palaeomagnetic data in north and west Europe, which shows upper Ordovician (Caradoc and Ashgill) magnetism with steep inclination, implying a new polar position, contrasting with earlier and later data with significantly lower inclinations (Piper, 1987). Palaeomagnetic data from China also shows uppermost Ordovician poles differed in position from those earlier and later (Wang Xiaofeng *et al.* 1983). If these proposed unusually high rates of continental movement are confirmed they could have a significant bearing on the growth and decay of the late Ordovician ice caps (Piper, 1987).

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